STRUCTURAL AND METAMORPHIC EVOLUTION OF A GNEISS TERRAIN IN THE NAMAQUA BELT NEAR ONSEEPKANS, SOUTH WEST AFRICA

by

D.J. TOOGOOD.

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ABSTRACT

A 4000 km^2 area of Precambrian rocks in the Namaqua belt has been examined and it is concluded the present structural and metamorphic framework is the result of a complex polydeformational and polymetamorphic evolution.

A major crustal break is present in the area and is represented by a northwest trending dextral shear zone - the Pofadder ZAHNCAFS. The zone of reorientation associated with this shear zone controls the geometry of the western part of the area. The shear zone varies in width from 20 to 40 km and the core contains a 2 km - 7 km wide belt of mylonites. Two sets of folds (D_5 and D_6) have been formed in the zone of reorientation in the northern block. The D_5 folds are northeast trending en-echelon structures up to 30 km long and the interference pattern produced by superimposed northwest trending D_6 folds has resulted in a series of crescent-shaped antiforms. These folds are not present in the southern block and this is thought to be due to a pressure shadow effect connected with the nearby Vioolsdrif complex. Components of both pure shear and simple shear were involved in the shear zone development and a minimum displacement of 85 km is indicated by the strain analysis. The shear zone developed under medium to high-grade metamorphic conditions and the mylonites were formed by a process of dislocation and recovery.

Harmonic and t_1/α analyses of mesoscopic folds associated with two older fold sets (D_3 and D_4) has shownthat certain fold attributes such as shape and 'tightness' are dependent on lithology and are virtually the same for all major structures examined. Mesoscopic folds in the Onseepkans area, therefore, cannot be regarded as having a unique style for different generations. Finite strain estimates show that D_3 folds in both the east and the west of the area represent a total shortening $\sqrt{\lambda_2/\lambda_1}$ = ,11 and $\sqrt{\lambda_2/\lambda_1}$ for D_4 folds varies between ,16 in the west and ,15 in the east.

Two major metamorphic events have affected the rocks. The first event (Kumian) produced granulite-grade metamorphites and was associated with the intrusion of norites and enderbites of the charnockitic suite. The Kumian is thought to have produced garnet and sillimanite in rocks of pelitic composition. Norites; enderbites, granulite-grade metamrophites and garnetiferous rocks of pelitic composition are absent from that part of the area south of the mylonite belt in the Pofadder ZAHNCAFS. The indicated P-T conditions for this event are 8 - 9 kb at 800°- 860°C.

The second event (Velloorian) was associated with the intrusion of two granitoids (Beenbreek megacrystic granite and Naros granitoid) and culminated in a period of widespread migmatisation. The Velloorian produced high-grade 'quartz + muscovite out' assemblages over most of the area and gave rise to the mantling of garnet by cordierite in rocks of pelitic composition and the retrogression of norites, enderbites and granulite-grade metamorphites. South of the mylonite belt in the Pofadder ZAHNCAFS 'quartz + muscovite in' assemblages define medium-grade metamorphites and rocks of pelitic composition contain sillimanite and prograde cordierite. The indicated P-T conditions for this event are estimated at 650° between 3.5 and 5 kb for the southern block and 650°- 700° between 5 and 7 kb for the northern block.

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Mineral Abbreviations used in Text

Al	Albite	Ky	Kyanite
Am	Amphibole	Mg	Magnetite
Ad	Andalusite	Mc	Microcline
Al	Almandine	Mu	Muscovite
An	Anorthite	Ol	Olivine
Bi	Biotite	Op	Orthopyroxene
Cc	Calcite	Or	Orthoclase
Cd	Cordierite	Pg	Plagioclase
Ch	Chlorite	Px	Pyroxene
Co	Corundum	Py	Pyrope
Cp	Clinopyroxene	Qz	Quartz
Di	Diopside	Sc	Scapolite
Ep	Epidote	Se	Sericite
Fo	Forsterite	Si	Sillimanite
Ga	Garnet	Sp	Spinel
Gr	Grandite	Tc	Talc
Hb	Hornblende	Tr	Tremolite
Hy	Hypersthene	V	Vapour (CO_2, H_2O)
Il	Ilmenite	Wo	Wollastonite
Kf	K-feldspar		

Symbols used in Structural Text

$\dot{e}$	strain rate
λ_1, λ_2, λ_3	principal quadratic extensions
X, Y, Z	strain ellipsoid axes
a.b.c	simple shear translation axes
ψ	angular shear
γ	shear strain ($\tan \psi$)
σ_1, σ_2, σ_3	principal stresses
μ	viscosity shear modulus
τ	shear stress
e_1, e_2, e_3	principal extensions

CHAPTER I

INTRODUCTION

The area described in this report lies in the extreme southeast corner of South West Africa. It is bounded by the Orange River in the south, the 19^{o} meridian in the west and the limit of the pre-Nama basement in the north. This is an area of approximately 5000 km^2 of which nearly 4000 km^2 is exposed rock.

The area forms part of the deeply incised valley of the Orange River, the only perennial river in the area. Rugged terrain with a relief of 600 m is typical in the immediate vicinity of the river (Orange River Mountains) but farther to the north extensive plains with scattered inselbergs are more characteristic. The northern limit of the area is marked by an escarpment formed by the undeformed Nama rocks whose height varies from about 250 m in the east where the escarpment borders the Orange River to about 20 m in the west where the escarpment is about 60 km from the river. This area is a desert receiving about 50 - 85 mm of rainfall a year. Most of this rainfall occurs in the summer months during violent thunderstorms. This section of the Orange River valley has the highest mean temperatures of any locality in southern Africa and summer screen temperatures of approximately 50^{o}C are not uncommon.

Vegetation is sparse except in the courses of the larger intermittent rivers such as the Keinab and Ham Rivers where tamarisk and thorn trees are fairly common. Thorn trees with a thick undergrowth line the banks of the Orange River. A detailed description of flora may be found in Beukes (1973). Several species of gazelle and antelope are present in the area as well as baboons, monkeys, leopard and jackal. Cobras, adders and scorpions are sometimes encountered.

There are no villages in the area. South of the river the small community at Onseepkans exists by irrigation farming on the flood plain of the Orange River. Cotton, lucerne, wheat and citrus fruits are grown. Elsewhere, most of the area is too arid to be used for anything except Karakul sheep farming.

1. *Previous work*

Apart from brief observations by Haughton and Frommurze (1936) there was no published information on the area at the start of the investigation. The only publication of note in the near vicinity was on the Riemvasmaak area some distance to the east (von Backström 1967) and the Pella area southwest of Onseepkans (Coetzee 1941). Outlines of the geology south of the river exist in Gevers et al. (1937) and Hugo (1969) both of whom were mainly concerned with the pegmatites.

Truswell (1970) includes a map of northern Cape Province and southern South West Africa in his book where rocks incorrectly described as 'Kheis System' are shown extending across the Orange River east of Onseepkans approximately where the Naros granitoid has now been mapped.

During the present investigation several other studies on adjacent areas appeared in print. Geological Survey 1:125 000 series maps along the southern bank of the Orange River extending from Goodhouse to Onseepkans appeared in 1973. This was, however, based on field work undertaken in the early 1940's. In 1974 unpublished D.Sc. theses on the Warmbad area (Beukes 1973) farther west and the Riemvasmaak area near Upington (Geringer 1973) became available (c.f. inset on geological map). The former area shares a common boundary with the present area under investigation, the latter a point of contact where the 20° meridian crosses the Orange River. Joubert (1974b) has published an outline of the geology of the Pofadder area to the south of Onseepkans.

2. *Regional geology*

The metamorphic rocks in the Onseepkans area appear contiguous with similar rocks found over large areas of Bushmanland, Namaqualand, northern Cape Province and southern South West Africa. They have variously been referred to as the Namaqualand gneisses (Joubert 1971), Namaqualand Granite Gneiss Massif (Martin 1965; Clifford et al. 1975a), Namaqua granite gneiss (Truswell 1970) and the Namaqua Metamorphic Complex (Blignault et al. 1974). They have also been referred to as forming part of the Kibaran Orogenic Zone (Holmes 1951), Sonama Crustal Province (Pretorius 1974; Cornell 1975), Namaqua-Natal Mobile Belt (Clifford 1970), Namaqua Mobile Belt (Vajner 1974) and the Namaqua Tectonic Province (Blignault et al. 1974).

In this report the term Namaqua-Natal Mobile Belt is retained for the belt as a whole and that section of it in the northern Cape Province and southern South West Africa is referred to as the Namaqua belt. The term is used for all metamorphic and associated igneous rocks underlying the Nama (Germs 1972) Gariep (Kröner 1974a) and Koras (du Toit 1965) Groups and formations of equivalent or younger age and, in the east, extending to the Kaapvaal Craton (Pretorius 1964). This includes the Vioolsdrif igneous complex, an area of mainly low to medium-grade metamorphic rocks derived from intrusive granitoids and lavas in the Goodhouse-Vioolsdrif area (Bertrand 1976; Blignault 1974; Reid 1974). It also includes those rocks described as part of the Kheis System (c.f. comment in section *1*). A sketch map of the region is shown in Fig. 1.

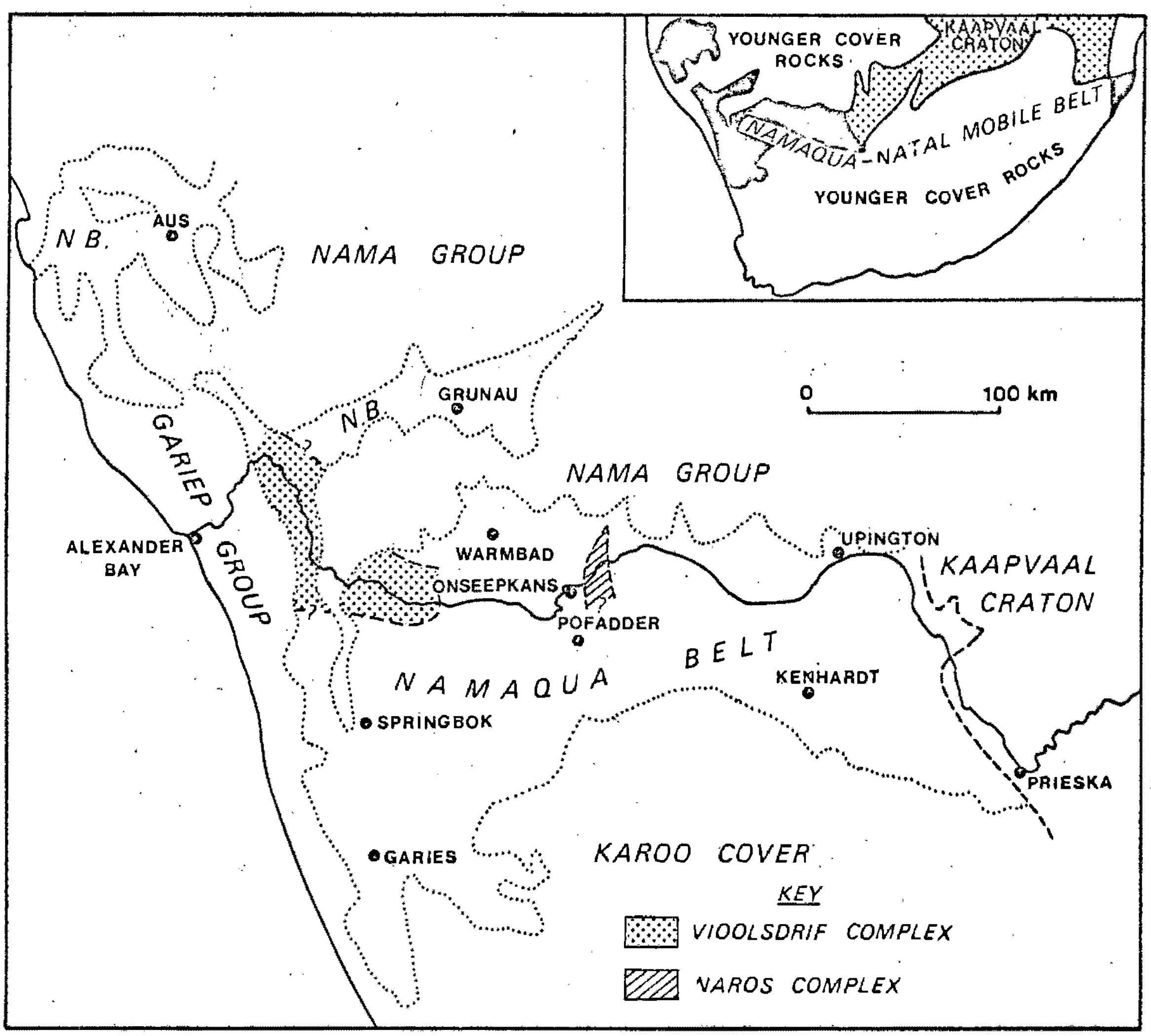


Figure 1. Sketch map of southern South West Africa and northern Cape Province showing the extent of the Namaqua belt (N.B.).

3. Present investigation

The aim of the present study was to provide a more detailed account of the metamorphic and structural evolution of a section of the Namaqua belt. Prior to this investigation only the results of one reconnaissance study covering 12 000 km^2 had appeared in print (Joubert 1971). Detailed investigations that had appeared dealt with specific problems such as the genesis of ore-bearing intrusives (Benedict et al. 1964) or radiometric age dates (Nicolaysen and Burger 1965). Other publications have already been mentioned in sections *1* and *2*.

For this purpose a geological map was made of the area which involved 13 months fieldwork. Mapping was done on 1:38 000 airphotos and then transferred to 1:100 000 preliminary topographic maps.

4. *Acknowledgments*

This study was initiated by the late Professor J. de Villiers to whom go special thanks. The project continued under Dr. A. Kröner whose supervision is gratefully acknowledged. Financial support came from the O'okiep Copper Company and the author is indebted to them for their assistance. The writer would also like to thank Dr. J. Gurney of the Geochemistry Department, University of Cape Town for making one garnet analysis. Dr. P. Fleming and Mr. C.J. Hartnady read parts of the manuscript and their comments are appreciated. This report would not have reached its present form without many discussions and advice from other members of the Precambrian Research Unit, in particular M.P.A. Jackson, C.J. Hartnady and Dr. R. Schultz. Thanks go to Miss P. Eloff who drew the maps and figures and Mrs. J. Macdonald who typed the report.

A special word of thanks go to Father R. Bientz and the Sisters of the Roman Catholic Mission, Onseepkans, whose unstinting hospitality was largely responsible for the successful completion of the fieldwork and whose friendship will always be valued. Finally, I would like to thank my wife, Heidi, for processing the photographs and for support throughout the study period.

CHAPTER II

LITHOSTRATIGRAPHY

This section describes the main lithostratigraphic characteristics of the rock types occurring in the area. It is not intended here to give a comprehensive description of the mineralogy since this is covered in Chapter III on metamorphism and would therefore involve unnecessary repetition. It is hoped that the reader will acquire a basic familiarity with the various formations to enable him to follow the metamorphic development of the complex as described in the next chapter.

In applying formation names the writer's aim was to comply with the recommendations of the I.U.G.S. Subcommission on Stratigraphic Nomenclature (Hedberg 1970). Above all the following statement was considered paramount: 'The proposed unit should be described and defined so clearly that any subsequent worker can without doubt recognise the same unit' (ibid., p.19).

The application of the I.U.G.S. system to metamorphic and igneous rocks is not entirely a straightforward process but the system remains the best available at present. Other classification systems in metamorphic terrains tend to be subjective and open to other interpretations, for example the terminology presented such as 'supracrustal', 'infrastructure', 'upper crustal' is poorly defined and equivocal. Such classifications are not compatible with modern trends which seek to offer an unambiguous frame of reference. The writer believes that even in highly complex areas of metamorphic and igneous rocks the classification should involve observable physical features rather than inferred geological history or mode of genesis (Hedberg op. cit., p.5).

With the increasing scale of mapping activity in the mobile belt there has been an unwelcome proliferation of formation names. It was found however that these terms could not be satisfactorily applied in the Onseepkans area, Beukes (1973) for example has a very extensive formation in the Warmbad area, the Umeis Formation or 'grey gneiss' which appears to be composed of many rock types mapped separately in the Onseepkans area.

Broadly speaking the formations described below may be divided into two groups, either pre-tectonic or syn-tectonic. The former group comprises a layered sequence of gneisses, the Austerlitz, Orangefall, Keimas, Pelgrimsrust, Jerusalem, Kambreek and Pella formations, which may represent a metasedimentary sequence. In the latter group occur all those rocks, granites, migmatites, etc., that were formed during the tectonic evolution of the mobile belt; these include the Beenbreek, Naros, Nautsis, Kum Kum, Grass River, Stolzenfels, Jericho and Skimmelberg formations. The layered pre-tectonic gneisses are collectively referred to in this text as the Onseepkans sequence.

The scale of the accompanying map (Annexure 1) does not permit narrow horizons of individual formations to be shown and rather than make the map illegible they have been omitted. Amphibolites of the Jericho formation are very abundant in some localities but they are only represented on the map where they help to define major structures in otherwise uniform rocks. The Orangefall and Pelgrimsrust formations are particularly prone to variation on a metre scale which cannot be effectively shown on the map and the Beenbreek granite is commonly full of deformed xenoliths and remnants of pre-tectonic gneisses. Pegmatites are very common throughout the area, but have been omitted from the map for the same reasons.

In the Warmbad area Beukes (1973, pp.107-108) has presented a structural-stratigraphic sequence which places quartzo-feldspathic gneiss (Houms Rivier Formation) at the base, mixed gneiss or 'grey gneiss' (Umeis Formation) containing 10 different rock types above and aluminium-rich gneisses (Arus Formation) at the top. This column was derived from a structural succession visible in large antiforms south of Warmbad which show extensive quartzo-feldspathic gneiss development in the fold cores. Since similar folds showing the same rock type in the cores are present in the Onseepkans area, it is important to establish whether this structural succession is viable on a regional scale.

On the farms Pelgrimsrust and Keimas a wide area of quartzo-feldspathic gneiss in an antiformal fold core appears to be structurally overlain by comparatively narrow bands of other rock types. An earlier fold has been mapped adjacent to this large structure and is refolded by it (Fig. 2). When examining the relationship between these two folds it becomes obvious that the quartzo-feldspathic horizon 'A' in the surrounding envelope of 'mixed gneiss' is almost certainly the equivalent of the large area of quartzo-feldspathic gneiss 'B' in the core of the late fold. It is equally apparent that the contact between the quartzo-feldspathic gneiss in fold core 'C' with the mixed gneiss envelope cannot be the same contact as that between fold core 'B' and its envelope. It is clear from this example that if relatively thin horizons of different gneisses are mapped as one formation the true relationship between individual units is lost. The apparent great thickness of quartzo-feldspathic gneiss in the Onseepkans-Warmbad area is partly due to its preservation in the cores of large-amplitude late folds. Since limb dips on these structures are commonly low, the formation occupies areas of several square kilometres although the actual horizon being folded need only be a few hundred metres thick.

The writer has concluded that no structural succession based on folds formed late in the tectonic development of the belt can be considered meaningful and the complexity of the deformation makes it difficult to arrive at a lithostratigraphic succession based on early folds. In areas where the early folds have been mapped without the interference of late structures (e.g. along the Ham River) it can be shown that the formations do not provide a clearly defined succession but are completely intercalated. The map legend, therefore,

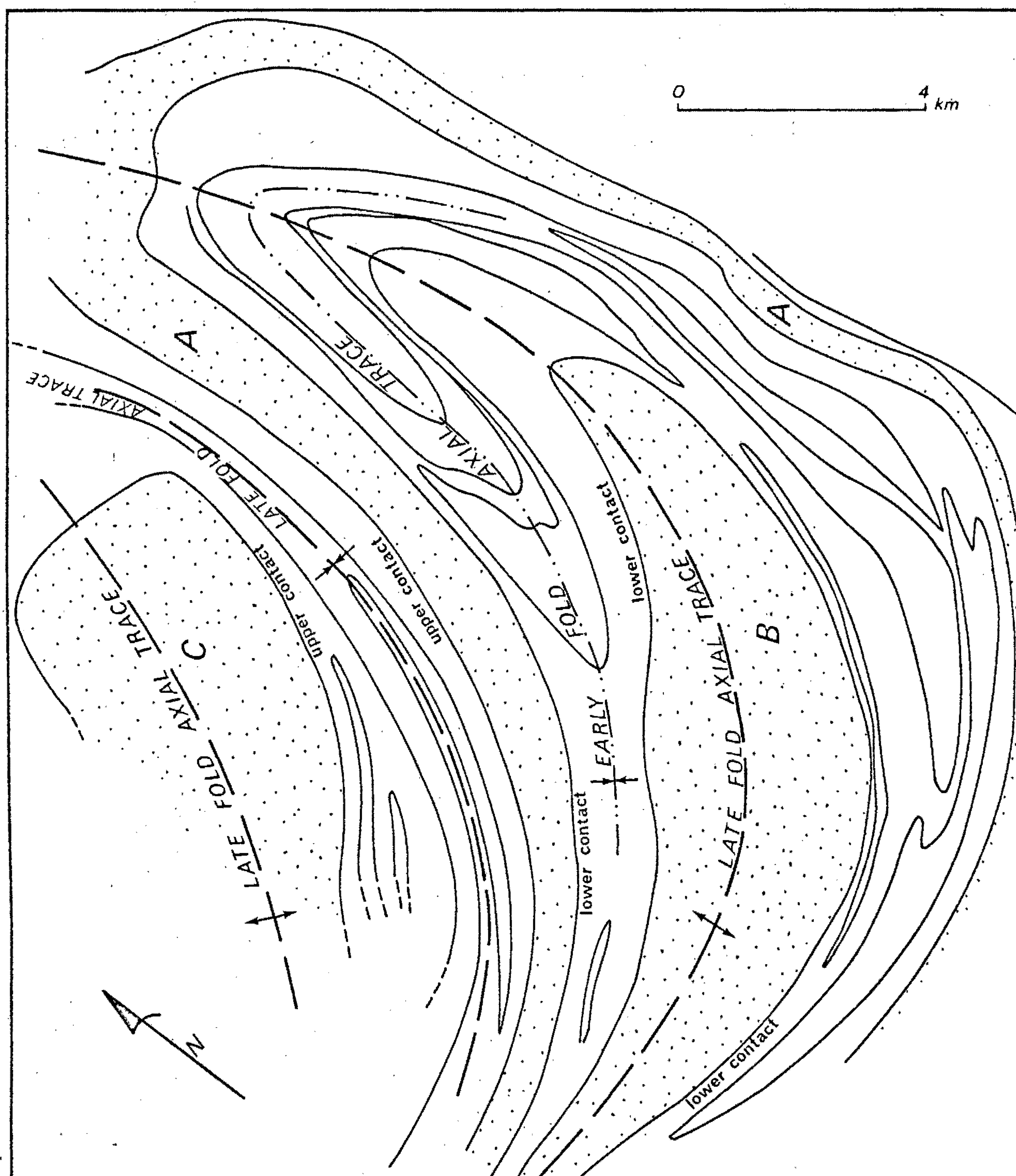


Figure 2. Simplified sketch of the geology west of Onseepkans. The quartzo-feldspathic horison 'A' is the equivalent of the quartzo-feldspathic gneiss in fold cores 'B' and 'C' thus demonstrating that the contacts between the fold cores and their envelopes are not the same.

cannot be taken in any way as representing a stratigraphic or structural succession in the Onseepkans sequence.

In the modal analyses listed below brackets () indicate minerals showing evidence of alteration. Fancy brackets {} indicate retrograde minerals. All analyses are estimated from charts prepared by Terry and Chilingar (1955).

A. THE PRE-TECTONIC GNEISSES - THE ONSEEPKANS SEQUENCE

1. *Austerlitz quartzo-feldspathic gneiss formation*

Rocks composed essentially of feldspar and quartz with a low content of mafic minerals (less than 10%) are widely distributed in the area and form one of the most common rock types in the Namaqua belt throughout the northern Cape Province and southern South West Africa. These rocks have been examined by several authors, notably Poldervaart and von Backström (1949) in the Kakamas area, Jansen (1960), Kröner (1968) and Pike (1959) in the Bitterfontein area. Benedict et al. (1964) in the Okiep area, Joubert (1971) in the northwestern Cape Province, Coetzee (1941) and Joubert (1974b) in the Pofadder area and Vajner (1974) in the Marydale area. The reader is referred to Kröner (1968) for a detailed discussion on the controversial origins proposed for rocks of this composition.

Various names have been used to describe these rocks, the most common being 'pink gneiss' but also leptite, leptynite, aplogneiss, granulite and biotite gneiss. The writer proposes that the term pink gneiss should not be used for lithostratigraphic classification. It originated as a field term describing the weathered appearance of certain rocks but its application throughout the Namaqua belt has led inevitably to the classification of rocks with diverse origins under a single heading. Leptite, leptynite and granulite are terms with a variety of definitions and in local geological publications these names have sometimes appeared without any definition. In general these rocks do not contain sufficient biotite to be called biotite gneiss. The writer favours the purely descriptive terms quartzo-feldspathic gneiss although the usage of the word gneiss for these often poorly foliated rocks is open to question. Winkler (1974) has suggested the name 'fels' for isotropic metamorphic rocks but this term has not become established.

The Austerlitz formation is composed of pink, grey or white quartzo-feldspathic gneisses which weather to a light brown colour. In the field they commonly give rise to smooth, upstanding, rounded outcrops but increasing mafic mineral content facilitates weathering thus producing a friable rock. In thin section they are seen to be composed of quartz, microcline, plagioclase and small amounts of biotite. Some rocks are composed essentially of quartz and plagioclase, usually andesine, but these are uncommon (see Table 1). Other minerals include muscovite, garnet, hornblende, sillimanite, sphene, epidote and chlorite. The mafic mineral content varies considerably. In some localities (e.g. on the farm Stolzenfels) the gneiss has a very low biotite content (<1%) but this rock type grades imperceptibly both across and along strike into horizons having a normal biotite content of between 2% and 8%. Sometimes the mafic mineral content varies in a seemingly erratic manner resulting in small irregular patches rich in biotite. Some localities show an abundance of small knots or ellipsoidal aggregates of sillimanite, quartz and muscovite up to 10 cm in length. Narrow horizons of red weathering schist rich in sillimanite are sometimes encountered in this formation.

TABLE 1.

Estimated modal analysis of rocks of the Austerlitz formation

Sample No.	Mu	Bi	Si	Ga	Kf	Pg	Qz
20		8			50	4	30
54					50	1	50
58		2			30	10	50
68		1			70	10	20
80		2				10	80
84		2			60	3	40
114						80	15
114A		4			70	1	30
187	3	5			70		30
199		8			50	20	20
216		(10)			20	30	40
271		10				50	30
287		2			20	30	50
325		(1)			50	(4)	50
380		10			20	50	20
1105					50		40
1136	5				30	2	60
1142	1	2			60		20
1142A	2	1			50		50
1191		5			60	1	40
1297						10	90
1263A	{1}	3			60		40
1275	{1}				60	2	40
1301		7			40	4	50
1306	2	1	5		30	5	50
1418						60	40
1141		1				30	70
1686	{1}	3				30	70
1481		(2)			60	5	30
1681		1		5		10	80
1817	{1}	1		1	80		20
25						40	80
56B			1	2	5	5	90
77		6			50		40

Sample No.	Mu	Bi	Si	Ga	Kf	Pg	Qz
794		3			40	40	20
798	{1}	(5)			80		20
689	{3}	1			40		60
784A					60		40
742			5	1	80		20
801	{4}				20	50	20
815A		1		1	40		60
824		5			40		60
828	{1}	8			40	(3)	50
832		7	1		70		20
834	{1}	8			20	(8)	60
868	{1}			2	50		50
870	{1}				50	15	40
872		4		9	30	(10)	50
874					40	(30)	30
878		1		2	90		40
893	{4}				10	10	80
900				2	80	5	20
744		5		1	80	2	15
905	{1}		2	1	70	10	20
916		6			70	3	20
917	{5}				80		20
917A	{5}	(3)			70	2	30
918		2			60	2	40
919	{1}	1			80		20
921	{2}	1		2	70	1	20
922			1	4	80		20
925					30	20	60
926	1	(7)		1	90	1	3
934		5				50	50
945		(8)			50		40
949		1			80	1	20
959	{1}	1			70	3	30
960				5	60	8	30
968		1		5	30	20	50
980				1	40		60
1062	{1}	1			20	10	70
756	{1}	5			20	10	60
783	{10}	(3)			8		80
911	{1}	6			70	1	20
926	{2}	(7)			80	1	2
708	{2}	(3)	1		80		20
937				6	50	10	40
2154	{1}				50	30	20
813	{1}		2		60		40
1130	1	1			40	5	60
1891	1	6			70	2	20
1934		7			50		40
199		8			50	20	40

Contacts between the Austerlitz formation are sharp except in two general cases. Increasing mafic mineral content gives rise to rocks classified as biotite gneisses which contain at least 10% biotite. In some localities it is not possible to define an exact contact and there are places where borderline cases had to be assigned to either the Austerlitz or the Orangefall formation. On the official Geological map of the Onseepkans area (1972), south of the Orange River, these mica rich horizons in 'pink gneiss' have been classified as 'paragneiss' but in a rather arbitary fashion that the writer was not able to verify. For example, where the axial trace of the 'Falls antiform' crosses the Orange River a perfectly symmetrical arrangement of rock types exists on each side of the axial trace which is not at all revealed by the disposition of the 'paragneiss' units shown on the official map.

Occasionally the Austerlitz formation occurs in contact with augen gneiss. Since this latter rock type originated by blastesis it is obvious that gradational contacts are the rule rather than the exception. Porphyroblasts of feldspar in the gneisses are fairly widespread, and where this fabric completely dominates the rock the name Grass River augen gneiss formation has been applied.

The quartzo-feldspathic gneisses provide little fabric data. There has not been sufficient ductility contrast within the gneisses to produce mesoscopic folds and a weak rodding is usually the only indication of large structures affecting the rock. Leucosomes which are fairly common in other formations are very poorly developed.

As mentioned above the origin of these gneisses has been subject to controversy, and since no chemical analyses are available from the area discussed here, the writer is not able to comment on most of the theories. The following observations, however, suggest that the Austerlitz formation originated as a psammite:

(i) narrow horizons of metaquartzites and, possibly, metapelites are fairly continuous within the formation.

(ii) at one locality on the farm Jericho cross bedding was found.

(iii) nowhere in the area was an intrusive contact seen between the Austerlitz formation and other rock types. In some localities a white garnetiferous aplite associated with the Beenbreek granite shows intrusive relationships. When deformed it is very similar in appearance to gneisses of the Austerlitz formation and it is shown on unpublished official maps as 'pink gneiss'.

2. *Orangefall biotite gneiss formation*

In this formation are included all the biotite gneisses present in the area. They are composed of biotite (>10%) quartz, microcline and plagioclase with lesser amounts of sillimanite, garnet and cordierite (Table 2). In the field the gneisses of this formation are light-to-dark grey in colour and commonly weather to a rusty brown colour.

The formation is lithologically variable and contains narrow schistose horizons rich in biotite, cordierite or sillimanite, fairly common biotite-poor horizons and minor hornblende gneiss. On the farms Ondermatjie and Duurdrift Sud, the formation is typically a banded sequence of these different members intercalated with quartzo-feldspathic gneisses of the Austerlitz formation and amphibolites of the Jericho formation. In the west the formation is represented by a more monotonous succession of biotite gneiss, but near Onseepkans it contains a profusion of zoned calc-silicate boudins (Plate 1). These boudins are the tectonically disrupted remains of formerly continuous layers.

Contacts with other rock types are sharp except, as previously noted, with the Austerlitz formation. Intercalated horizons of augen gneiss occur in the west and show gradational contacts with biotite gneisses.

The biotite gneisses provide a wealth of fabric data in the form of lineations and folds. The rocks are invariably migmatised to some degree; southwest of Onseepkans the neosome is particularly abundant.

Biotite gneisses are widely distributed throughout the Namaqua belt and are subordinate only to quartzo-feldspathic gneisses in the scale of their development. In Namaqualand they have been described as semi-pelites (Joubert 1971) and in the Warmbad area as grey gneiss and granite gneiss (Beukes 1973).

TABLE 2.

Estimated modal analysis of rocks of the Orangefall formation

Sample No.	Mu	Bi	Si	Ga	Cd	Kf	Pg	Qz
289		12		1			20	70
1181		20	2			65		15
1273		15		3			5	80
1437		20		2			10	70
1824	{2}	10				60	1	30
1955		(15)				15	1	50
2131		15	1			30	5	50
299		10		1			20	60
1232	7	12	5			1		70
964		20		2			30	50
966		15					20	60
2386		20		5		10	5	60
2086		(15)						40
691		20					4	80
724	1	20	4	1		1		70
813A		10	1	2	1	5		90
852		10	2			20	5	60
860		10	1			20		70

Sample No.	Mu	Bi	Si	Ga	Cd	Kf	Pg	Qz
877	{2}	20						70
921		20				5	(40)	30
951		10		1		1	10	80
576		15	1			60	2	20
91	1	15					20	60

3. *Jerusalem cordierite-sillimanite-garnet gneiss formation*

Included here are all those rocks which contain sillimanite, garnet or cordierite in significant amounts (Table 3). The Jerusalem formation is the equivalent of the pelitic rocks or aluminous gneisses of Joubert (1971, p.147-164; 1975, p.340) and the Arus Formation of Beukes (1973, pp.147-164) which is also described as a metapelite. The term pelite in metamorphic petrology has been applied to rocks with a somewhat atypical pelitic composition (Leake 1958, Chinner 1961). This is because a sedimentary origin for these rocks can only be inferred when the Al_2O_3 content is so high that any possible igneous parentage can be discounted. Thus rocks with a typical pelitic component of 20% Al_2O_3 (often described as metapelites) are not indubitably paragneisses. From chemical analyses of the Arus Formation, it appears that a high (∿30%) Al_2O_3 content requires the rock to contain more than 90% combined micas, sillimanite, cordierite and garnet (Beukes 1973, pp.153 and 159 samples G.B.W. 98 and 302). This percentage is seldom realised in rocks belonging to the Jerusalem formation and, strictly speaking, the rocks of the formation may not, therefore, be referred to as metapelites. Most commonly the gneisses of this formation are fine grained and dark grey or dull green in colour. They are poorly foliated and may give the impression of being an igneous rock. Along the Ham River irregular patches of pink or amber feldspar are present but usually these rocks are never strongly migmatised. Narrow horizons of these gneisses in the Austerlitz formation are sometimes very coarse grained with bladed sillimanite crystals 15 cm in length. The Jerusalem gneiss is not a very common formation in the Onseepkans sequence and it is normally encountered as thin horizons in the biotite gneisses or quartzo-feldspathic gneisses. The most extensive exposures occur on the farm Jerusalem.

TABLE 3

Estimated modal analysis of rocks of the Jerusalem formation

Sample No.	Mu	Bi	Si	Cd	Ga	Kf	Pg	Qz
757	{2}	15	8		1	10		60
832A		30	20	4	15			30
837A		20	5	5	3	2	5	60

Sample No.	Mu	Bi	Si	Cd	Ga	Kf	Pg	Qz
838	{1}	10	4	4	5		5	70
914		5	10	10	8			60
890		5	5	10	5	1	5	70
845		15	8	10	10		5	50
1992		10	10	4	15	8	12	50
2567		15	10	4	7	3	15	50
835		10	4	1	2	30	3	50
815			7		15			60
1427		20	5	3	3			70
1403A		10	6	1	5			80
1479		8	10			50		30
73A		15	30	1	2	10		30
1509		8	7	1	8		8	70
1649		15	6	15	2		1	60
1428		15	20	5				60
36		15	15		10			60
13	{8}	15	7	4				70
1004			5	4	5	10		50
2011		10	5	3		40	2	90
2202		(5)	1	10		50	1	30
2205		7	5	4	5	30		50
2020A		30	5		1	10		40
2204		20	3	7		30	10	40
56A		1	4	2	15	40		50
1460F		20	8		5		4	60
1464		15	10	2	8		3	60
376	2	10	3	4				70

4. *Kambreek quartz-muscovite schist formation*

Although the schist of this formation are very similar in composition to the Jerusalem gneisses they are sufficiently distinctive in the field to warrant a separate description. They are composed of biotite (usually very pale), muscovite, sillimanite, quartz, cordierite, microcline and plagioclase (Table 4). This paragenesis imparts a very lustrous appearance to the rock. The schists occur as horizons a few hundred metres thick within the Austerlitz formation and are almost exclusively limited to the area south of the Pofadder Lineament.

TABLE 4.

Estimated modal analysis of rocks of the Kambreek formation

Sample No.	Mu	Bi	Cd	Sill	Kf	Pg	Qz
212		(20)			10	30	40
1139		(2)				10	20
1141	90						
1149	20	(15)					70
1150	3	(10)	3	5			70
1155B	5	(8)	4				40
1172	20	(3)	2	5	1		60
1176	3	(8)	5				60
1188A	15		8				40
1213	7	(10)		2			70
1216		(15)			60		15
1218	15	(8)	7				50
184	2	30		7			60
269	2	30	5	6			70
1121A	10	10			30		50
1128	5	15			30		10
1137A	20	7	1	2			70
1148	2	15		2	10		50
1200	5	20	1	10			70
273	5	20	6	5	1		60
1205	5	20		20			60
1151	6	25		10			60

5. *Pella metaquartzite formation*

Metaquartzites are quite uncommon in the area as a whole. They are mostly found south of the Pofadder Lineament (see map) and are one of two rock types (distinct from metamorphic parageneses) whose distribution appears to be controlled by displacement on this lineament. The thickest horizon encountered which was in the hinge zone of a fold was only 4 m in width. Very narrow bands of metaquartzite are also found in the Kambreek formation but it has not been possible to show all of them on the accompanying geological map.

These metaquartzites are the rather insignificant representatives of a very thick succession found at Pella immediately south of the Orange River. The Pella metaquartzites have been correlated by Coetzee (1941) with the Kaaien quartzites of the Upington area and Vajner's (1974) discovery of probable Kaaien quartzites in the eastern Namaqua belt lends support to this correlation.

In the field the metaquartzites are grey or white in colour with some horizons showing a fine banding. Biotite, epidote, muscovite, chlorite and plagioclase have been identified in thin section in amounts of 1% or less. North of the Pofadder Lineament a single 1,5 m thick horizon persists for some distance along strike on the farms Orangefall and Keimasmond. On the farm Vaaldoorn some poorly exposed quartz-rich rocks may be quartzites of this formation. In hand specimen these rocks have a strong brown colouration and often contain up to 20% opaque minerals.

6. *Pelgrimsrust hornblende gneiss and schist formation*

Extensive outcrops of hornblende schists and gneisses of the Pelgrimsrust formation occur west of Onseepkans. The deformation in this area is particularly complex and it is difficult to determine whether several stratigraphic horizons are involved or merely one horizon duplicated by folding. These rocks are poorly represented in the east which suggests that their true thickness may be comparatively small.

The significant feature of all the rocks making up this formation is the presence of hornblende as a major constituent. There is some variation in the appearance of individual horizons but the most common rock type is a dark-grey hornblende gneiss. Some horizons consist essentially of biotite-hornblende gneisses while in other cases hornblende appears as prominent euhedral crystals in a leucocratic gneiss. Horizons of chlorite and epidote schists are often encountered in this formation. The rocks of the Pelgrimsrust formation are essentially composed of hornblende, biotite, plagioclase and quartz with the addition of pyroxene, sphene, epidote and microcline (Table 5).

TABLE 5.

Estimated modal analysis of rocks of the Pelgrimsrust formation

Sample No.	Bi	Hb	Px	Sn	Ep	Pg	Kf	Qz
60		15				60	5	20
74	3	40				50		10
82	15	8				20	40	15
86		20		3		70		10
110	15	7		2		30		8
123		60	2			30		8
134		25		3		40	5	30
156	(10)	5			1	(60)		20
251	20	10				50	5	20
253		30				(60)		10
166	10	10		1		(30)	15	8
310		30				40	10	30
353		30				(50)		15
377	15	10		2		10	40	20
1132B	2	30						
1133	(19)	20			3	(60)		7
1157	(15)	10		1	2	(50)	10	15
1261		70				(30)		
1318		10		1	20	50		30
1319		10				10		80
1416	1	50				40		8
1620		20				(50)		30
1917	(15)	5				(70)		10

The rocks of this formation are intercalated with quartzo-feldspathic, biotite and calc-silicate gneisses. Some horizons of hornblende gneiss are indistinguishable from amphibolites of the Jericho formation but because of the ambiguity involved, these have been included in the Pelgrimsrust formation on the geological map.

There are no indications in the Onseepkans area as to the origin of these rocks. Farther to the west Beukes (1973, pp.141 and 143) has found marbles associated with hornblende gneisses and suggested a metasedimentary origin for this sequence. A hornblende agmatite is found as a several hundred metre-thick horizon north of the Pofadder Lineament and as narrower horizons elsewhere in the Pelgrimsrust formation and is composed of angular fragments of hornblende rich rock in a leucocratic gneiss (Plate 2). It is not certain how the agmatite formed but it probably results from the partial migmatisation of the Pelgrimsrust formation (Mehnert 1968, pp.8-9).

7. *Keimas calc-silicate gneiss formation*

Narrow, usually discontinuous, bands of calc-silicate gneiss are found intercalated with the Austerlitz, Orangefall, and Pelgrimsrust formations. They are, however, virtually restricted to the western half of the area. They are usually light grey in colour but weathered surfaces are often black. They typically give rise to narrow resistant ridges in other formations, usually have a waxy lustre and produce a resonant sound when struck. Sometimes the mafic constituents are scattered through the rock but more often they occur in thin layers giving the rock a finely banded appearance. In the Orangefall biotite gneisses these rocks occur as lenses and boudins with a marked zoning.

Under the microscope they are seen to be composed of clinopyroxene, hornblende, plagioclase and quartz with smaller amounts of microcline, sphene, epidote, scapolite, tremolite and garnet (Table 6). Free carbonate is only found in trace amounts.

TABLE 6.

Estimated modal analysis of rocks of the Keimas formation

Sample No.	Am	Cp	Sn	Ga	Pg	Kf	Qz
83A	3	10	1		8	5	70
170A		7	1		30		60
171A	2		1		30		60
1271A		10			15		70
1304	10		1		30	60	1
1308		12	2		80	5	
1455A		8		25	50	7	

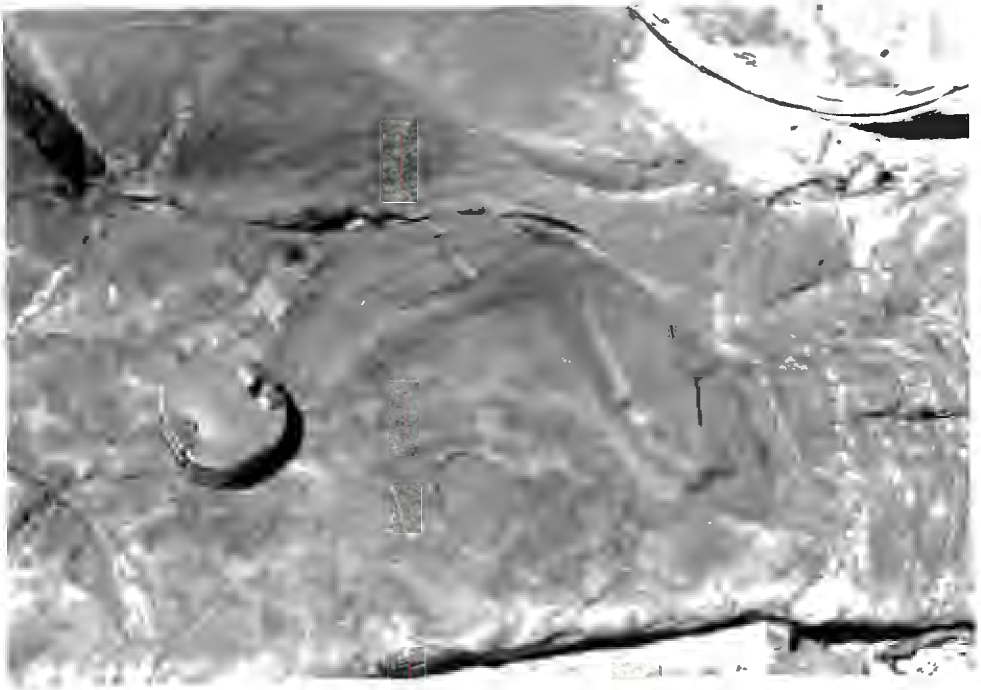


Plate 1. Zoned calc-silicate boudin in biotite gneiss (Orangefall formation). Keimasmond Farm.

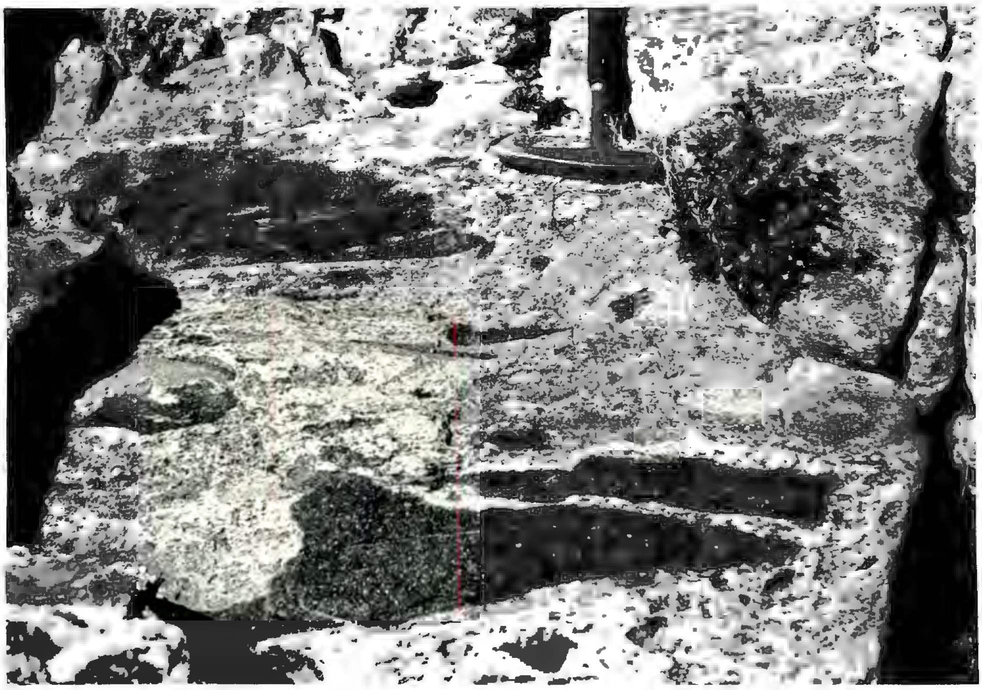


Plate 2. Hornblende agmatite composed of angular hornblende-rich fragments in a leucocratic rock. Orangefall Farm.

Sample No.	Am	Cp	Sn	Ga	Pg	Kf	Qz
1636	5	10	2		50		40
1683A	15	4			60		15
77A	20	15			40		20
109	20	10	2		15	15	30
1303	15	8	1		40		40
1388	30	30	1		10		15 + Sc
99A	5	20	2	10	20		40

These rocks are described as granulites and calc-silicates in the Warmbad area (Beukes 1973, pp.143-145), calcareous rocks in Namaqualand (Joubert 1971, p.45) and calc-silicates in the Pofadder area (Joubert 1974b). Most authors have interpreted them as metasediments. The writer prefers to restrict the term granulite to granulite-grade metamorphic rocks in keeping with recent views on the subject (Mehnert et al. 1972; Winkler 1974).

B. THE SYNTECTONIC GROUP

1. *Beenbreek megacrystic granite formation*

The Beenbreek megacrystic granite is the most widely distributed granitoid in the area. It is found associated with nearly all other formations, often in a very intricate manner. It is not found south of the Pofadder Lineament, however, and the equivalent of this granite in the Warmbad area (Beukes 1973, pp.202-220) is likewise restricted to the northern block. In composition it falls in the granite field (Fig. 3) and the most conspicuous feature of its mineralogy is the presence of large (up to 20cm) microcline and plagioclase megacrysts (Plate 3). Present in the matrix are garnet, biotite, sillimanite and sometimes cordierite and muscovite, together with plagioclase, microcline and quartz (Table 7).

TABLE 7.

Estimated modal analysis of rocks of the Beenbreek formation

Sample No.	Mu	Bi	Ga	Si	Cd	Kf	Pg	Qz
110A		20	2			50	6	30
293		25				20	30	30
296		15				40	15	30
327		15				30	20	40
368		15	3	5	1	5	40	40

Sample No.	Mu	Bi	Ga	Si	Cd	Kf	Pg	Qz
451	{1}	15	3			40	10	30
829		15	10	3	1	30	10	30
872		8	5			50	5	30
906	{1}	20				5	30	40
908		15	5			10	30	40
920		20	1			15	40	30
926	{2}	15	1			60	5	20
957		30				15	40	20
971		15				40	15	30
2553		20	1			20	20	40

In the east, especially on the farms Ondermatje and Jerusalem, the granite is intruded by aplite. The aplite is white in colour and is composed of quartz, microcline, plagioclase and garnet, with very minor biotite. In some localities blocks of granite appear as xenoliths in the aplite (Plate 4). This aplitic material has not been found in the west of the area and none is reported in association with the equivalent of the Beenbreek granite in the Warmbad area (Beukes op. cit.).

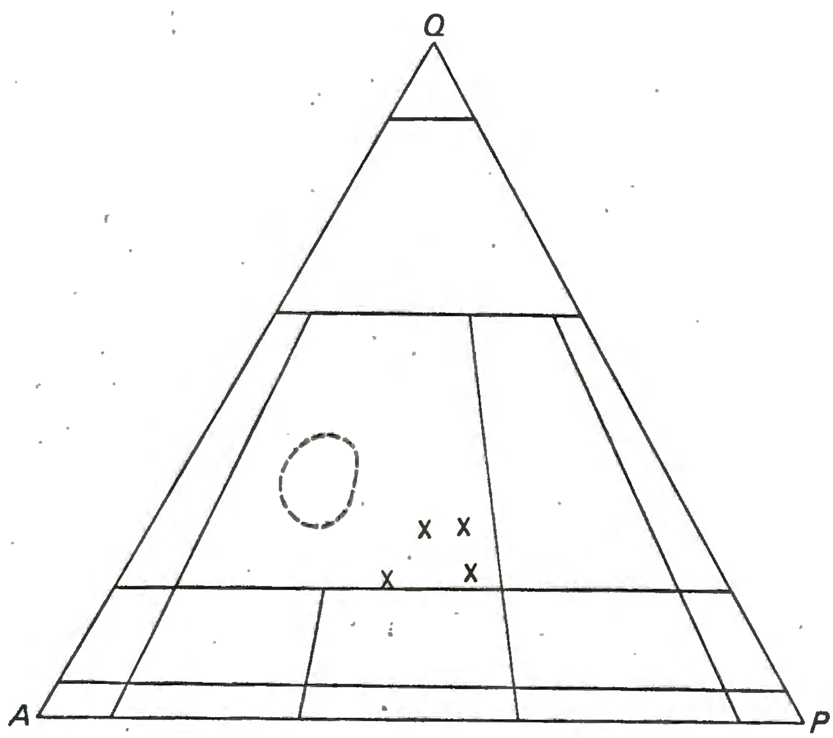


Figure 3. Composition of the Beenbreek megacrystic granite plotted on a classification diagram taken from Streckeisen (1973). Field outlined represents the composition of the Eendorn granite in the Warmbad area (Beukes 1973, p.214).

In most cases contacts between the Beenbreek granite and other rock types are sharp but no chilled margins have been seen. Clear intrusive relationships exist with rocks of the Austerlitz, Jerusalem, Stolzenfels and Kum Kum formations, and all but the first are found as xenoliths in the granite (Plate 5). However, in cases where the contacts are not sharp, megacrysts of feldspar can be seen forming in the adjacent country rock (Plate 6). Contacts of this nature provide evidence for a genetic relationship between the Beenbreek megacrystic granite and the widespread blastesis in the Onseepkans sequence.

Invariably the megacrystic granite has been foliated to some degree and is now represented by an augen gneiss. In some cases, it grades completely into garnet-biotite gneiss as a result of this deformation and no accurate contact with other rock types can be mapped. This problem is not significant when the area is taken as a whole since the Orangefall biotite gneiss formation can usually be distinguished by its variable composition and the presence of calc-silicate boudins. Where the granite contains aplite, deformation has produced an augen gneiss intercalated with garnetiferous-quartzo-feldspathic gneisses that sometimes appear very similar to rocks of the Austerlitz formation.

One outcrop of megacrystic granite close to the Udabis River on the farm Naros was found to contain 5% orthopyroxene. Field relations make it unclear whether the rock belongs to the Beenbreek formation or whether it forms part of the Naros granitoid. This rock can be classified as a charnockite but apart from the presence of hypersthene it is identical to the Beenbreek megacrystic granite.

Charnockites have been reported from the area northwest of Upington by Geringer (1973) but the present data are insufficient to suggest a correlation since it is possible, for example, that the hypersthene is a refractory phase resulting from assimilation of high-grade gneisses.

The Beenbreek megacrystic granite can unquestionably be correlated with the Eendorn granite in the Warmbad area (Beukes 1973, pp.202-220) and the Backrivier granite northwest of the Upington area (Geringer 1973, pp.98-107). It has many similarities with the Tsirub gneiss of the Aus area (Jackson 1976) as well as the Nababeep gneiss and the porphyroblastic granite gneiss described in Namaqualand and Bushmanland (Joubert 1971, pp.16-18 and 51-53; 1974b, p.340). Both the latter formations are reported to contain leptites or leucocratic bands which may be the local equivalent of the deformed aplites seen in the Beenbreek granite. Joubert (1971, p.51) noted that the porphyroblastic granite gneiss has definite intrusive relationships with the paragneisses and believes the intrusion is a northerly representative of the Nababeep gneiss (op. cit. p.14).

2. *Grass River augen gneiss formation*

There are a variety of microcline-plagioclase augen gneisses in the area and these may be categorised under four main headings

(i) augen gneisses resulting from the penetrative deformation of the Beenbreek megacrystic granite.

Plate 3. Beenbreek megacrystic granite. Beenbreek Farm.

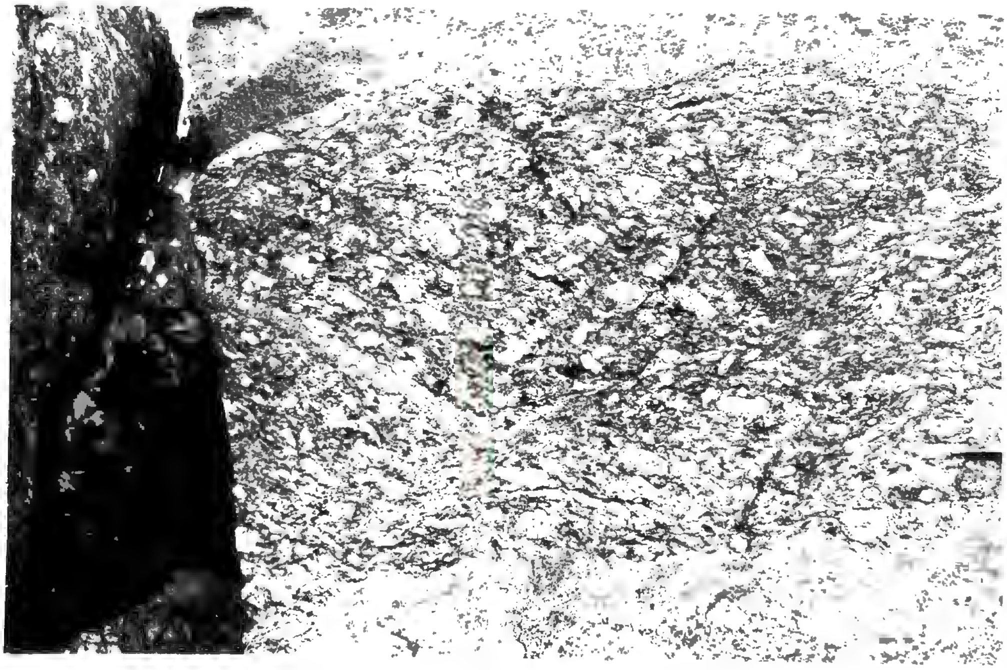


Plate 4. Blocks of Beenbreek megacrystic granite surrounded by garnetiferous aplite. Ondermatje Farm.

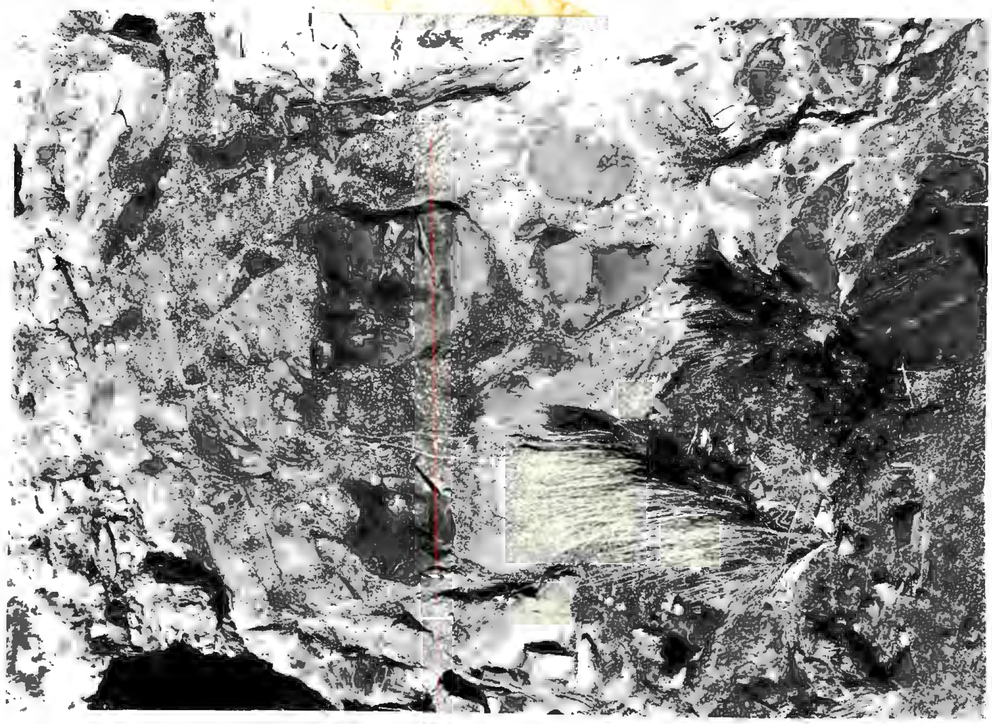


Plate 5. Xenoliths of mafic granolite (Kum Kum formation) and garnet-sillimanite gneiss (Jerusalem Formation) in Beenbreek megacrystic granite . Keimas Farm.

Plate 6. Contact between Beenbreek megacrystic granite and biotite gneiss (Orangefall formation) showing the development of porphyroblasts in the gneiss. Khaais Farm.

(ii) augen gneisses possibly resulting from the deformation of other, unrecognised, megacrystic granites.

(iii) augen gneisses resulting from blastesis.

(iv) augen gneisses resulting from the deformation of rocks in linear high strain belts (shear zones).

The last category is fairly simple to recognise and will be discussed in a later section. The decision as to which rocks of the former three categories make up the most of the Grass River formation remains one of the most intransigent problems encountered.

Blastesis is known to have affected the Onseepkans sequence since porphyroblasts can be seen in country rocks adjacent to the contact with megacrystic granite (Plate 6). The fact that augen can be seen in a wide variety of rock types including hornblende, biotite and quartzo-feldspathic gneisses (Plate 7 and Beukes 1973, Fig. 56) is characteristic of blastites and mitigates against an origin through deformation of a megacrystic intrusive. The Grass River formation on the farms Ondermatje and parts of Duurdrift Sud is thought to have been derived by blastesis.

Augen gneisses derived from the Beenbreek granite can usually be recognised by the presence of intrusive contacts, non-foliated areas, xenoliths and aplite material. These augen gneisses are therefore included with the Beenbreek granite on the map. Where there has been any doubt the rocks have been placed in the Grass River formation and the augen gneisses of the lower Ham River and around Onseepkans possibly belong to this category.

Augen gneiss on the farm Khaais has a marked igneous appearance since it displays a uniform composition and has no penetrative foliation. In the east of this farm the augen gneiss contains inclusions of Beenbreek granite but no definite intrusive contacts were found.

Farther to the west, on the farms Velloorsdrift and Keimasmond, the rocks of the Grass River formation have been strongly refoliated and the augen now define fold axial planes. The augen formed during this second cycle of deformation are smaller and it was impossible to determine whether they were derived from megacrystic granite or blastites.

TABLE 8

Estimated modal analysis of rocks of the Grass River formation

Sample No.	Mu	Bi	Ga	Si	Hb	Kf	Pg	Qz
371		20			5	10	30	30
426		20					30	40
841	{1}	30				60	4	20
932		15			2	30	10	40
936		10	1			60	5	20

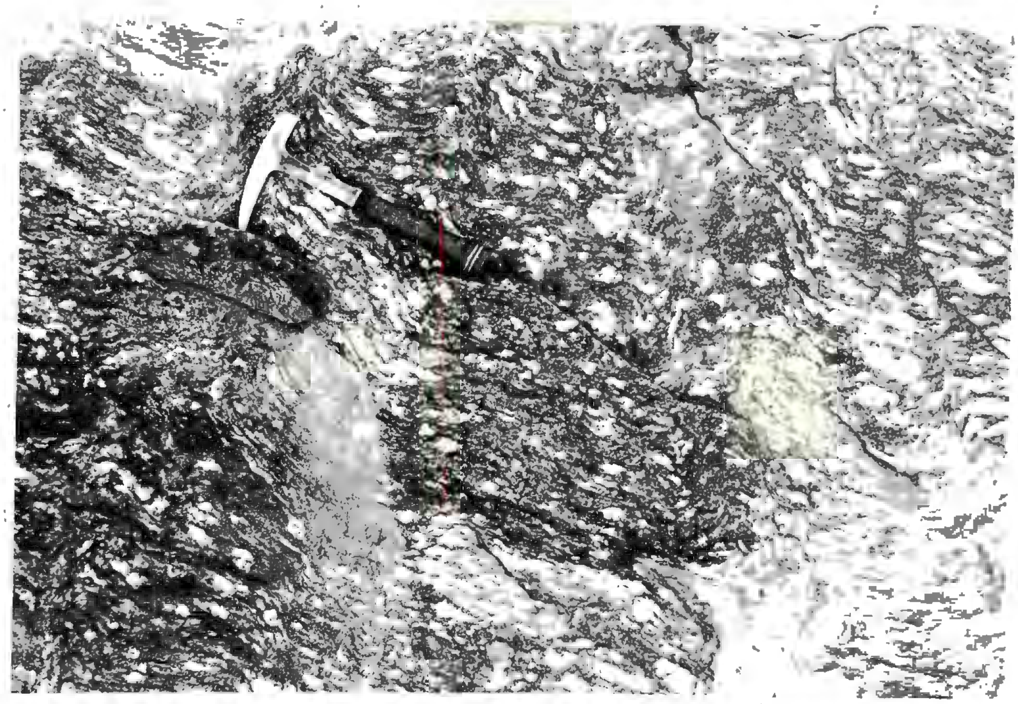


Plate 7. Feldspar porphyroblasts in marginal amphibolitised zones of the Stolzenfels norite and enderbite formation. Keimasmond Farm.

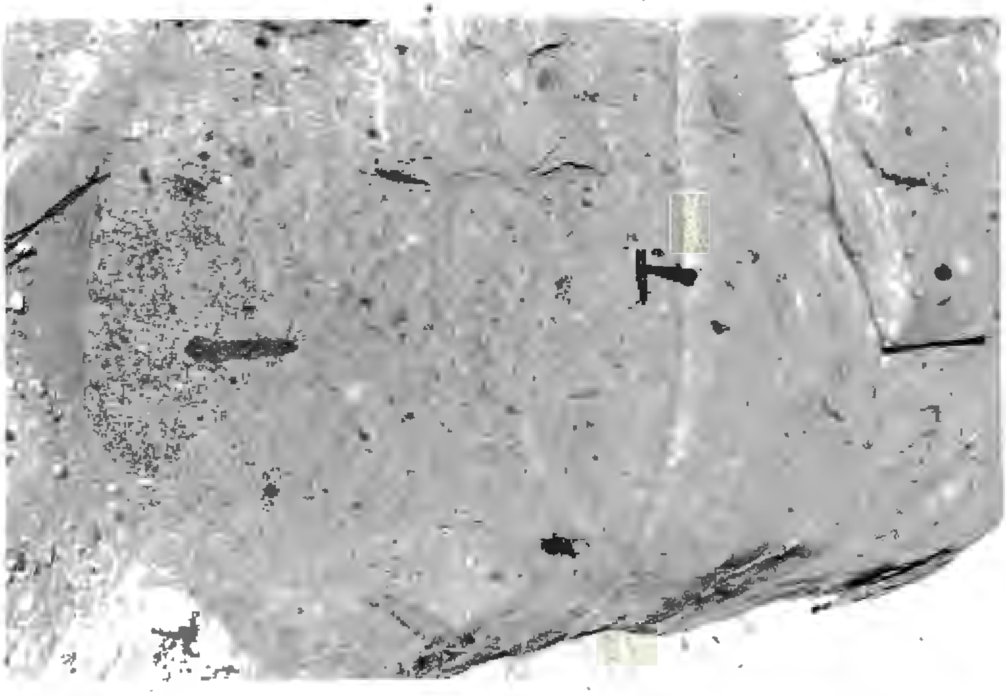


Plate 8. Naros granitoid with mafic inclusions. Ondermatje Farm.

Sample No.	Mu	Bi	Ga	Si	Hb	Kf	Pg	Qz
942		20				30	30	20
960		15			6	40	10	30
999		20				30	15	30
1484		15	1			5	20	6
1751	1	20					20	60
2071		20				40	10	30
2217		15				50	30	20
551		10	1				30	60
973		20			8	20	20	30

Beukes (1973, pp.135-139 described porphyroblastic biotite gneisses in the Warmbad area which are probably the equivalent of the Grass River formation. However, in at least one locality at the western boundary of the study area the porphyroblastic biotite gneiss was seen to represent deformed Beenbreek megacrystic granite. Joubert (1971, pp.47-49) states that the porphyroblastic gneisses mapped in Namaqualand are usually isotropic and in places intrude paragneisses.

3. *Naros granitoid formation*

A large granitoid intrusion straddles the Orange River approximately at its confluence with the Ham River. The granitoid has an hypidiomorphic granular texture and its essential constituents are hornblende, biotite, plagioclase, quartz and microcline. It varies in composition between granite and quartz- syenite (Streckeisen 1973) as can be seen from 4 samples plotted in Fig. 4. The fresh rock has a dark-grey colour and weathers a dull-brown colour, it typically produces a topography of low conical hills and loose rounded boulders.

The Naros formation contains an abundance of non-foliated mafic inclusions (Plate 8) and sometimes small leucocratic inclusions. The mafic inclusions are commonly ellipsoidal or cigar-shaped. The main intrusion is surrounded by a wide zone in which smaller stocks and sills of granite are intimately associated with the rocks of other formations. This zone is shown on the accompanying map as the Naros complex. Unequivocal intrusive relationships exist with respect to deformed Beenbreek granite and the Austerlitz formation (Plate 9).

A weak foliation can be seen at many localities and certain marginal zones are strongly foliated with the mafic inclusions drawn out and flattened (Plate 10). The development of the foliation is associated with the main period of migmatisation and is axial planar on folds defined by neosome bands.

Plate 9. Xenoliths of foliated Beenbreek megacrystic granite in Naros granitoid. Ondermatje Farm.

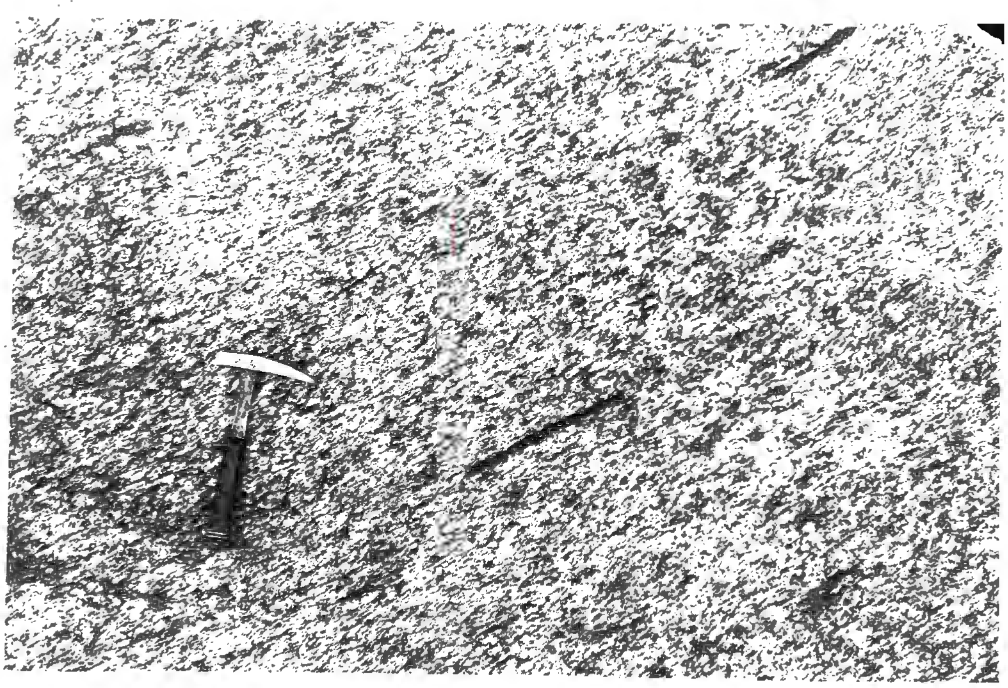


Plate 10. Foliated Naros granitoid with flattened mafic inclusions. Beenbreek Farm.

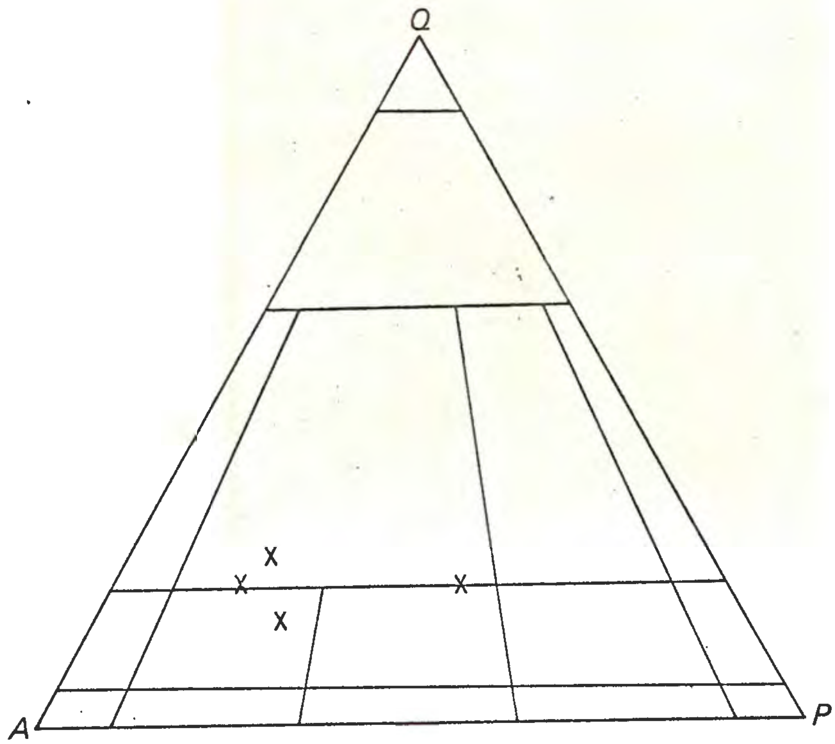


Figure 4. Composition of the Naros granitoid plotted on a classification diagram taken from Streckeisen (1973).

4. Nautsis alkali-feldspar granite formation

An igneous origin for this rock of granitic appearance can only be tentatively suggested since no intrusive contacts have been found and it contains no xenoliths. In many respects it resembles the quartzo-feldspathic gneisses but the complete lack of lithological variation and the absence of a fabric in the main body is not typical of the Austerlitz formation and on this basis it is classified as an 'igneous-looking' rock.

The granite forms the mountain Rooiberg on the farm Nautsis. Both weathered and fresh specimens are pink in colour and generally massive. A sill of this granite extends to the northeast where it now forms macroscopic fold cores. In this locality, and in some marginal areas, the rock is foliated and weakly migmatised.

Three specimens are plotted on a classification diagram in Fig. 5 and all fall in the alkali-feldspar granite field (Streckeisen 1973).

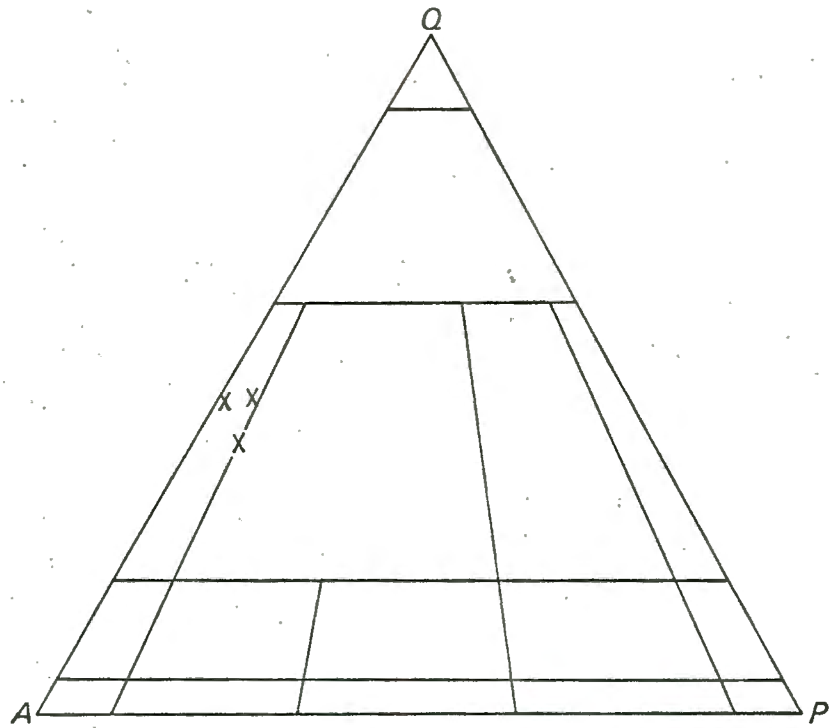


Figure 5. Composition of the Nautsis alkali-feldspar granite plotted on a classification diagram taken from Streckeisen (1973).

The rock has a hypidiomorphic granular texture. The plagioclase (An_{30}) is subhedral or anhedral and the K-feldspar consists of microcline-microperthite, mainly flame and ribbon varieties. Biotite (<5%) is largely altered to penninite and quartz shows weak strain extinction. There is no evidence of blastesis and no pegmatite is present. Two narrow amphibolite bands were found at one locality. The minerology of the granite is very uniform and the presence of microcline-microperthite is unusual only being encountered in significant amounts outside the granite in mylonites. These features also suggest that the Nautsis granite is not part of the Austerlitz quartzo-feldspathic gneiss formation.

5. *Stolzenfels norite and enderbite formation*

Four large and several small basic bodies are present in the area varying in composition from norite to enderbite. These bodies usually form rugged isolated hills and are easily recognised by their black weathered surfaces. They are composed of augite, hypersthene and plagioclase with the addition of biotite and quartz. Hornblende commonly occurs as an alteration product

(Table 9) but in some thin sections was also seen as a primary mineral.

TABLE 9.

Estimated modal analysis of rocks of the Stolzenfels formation

Sample No.	Bi	Cp	Op	Hb	Pg	Qz
33		40	10	{40}		
1459	2		30	{15}	50	
1463			15	10	70	1
1470	6	8	15		60	1
119	8	20	10		60	3
1365A	10			{20}	60	
1365B	15			{20}	60	
1424	15	5		{15}	50	
1374				{50}	50	
1413A	4		10	{20}	60	6
1432	7	(5)			50	3
1445	7	10	15	{10}	60	
1435	20	40			40	
1322		2		{30}	50	2
610	10	2	8	{15}	40	30
843	7			{40}	50	8
981		10	10	{40}	25	5
977	15	10	5		50	25
604	5		12	{2}	55	25
721	10		10	{4}	60	20
697	7	7	15		70	8
769		5	30		50	15
2571	1	15	20		60	2
2573	10	7	18		60	2
2566	15		5	{10}	60	7
1566A	30			{5}	50	20
1540		20	40		15	
366		8	10	{30}	40	15
1764			15	{30}	40	15
1821		20		{40}	30	10
2575	5	4	30		60	5
2568	5	6	6	{40}	40	1

The mineralogy and appearance of these rocks shows that they are mafic representatives of the charnockitic rock suite (see Chapter III) and they are important indicators of the metamorphic development of the mobile belt. The mafic bodies are generally structureless except for the sporadic appearance of a foliation. They are found in contact with rocks of the Orangefall, Austerlitz and Beenbreek formations. Contact zones are usually retrograded and foliated, completely destroying the original relationship with the country rock. Noritoids or rocks with an equivalent composition are found throughout the Namaqua belt. In the Nuwerus area they are described as norites by Brink (1950), in Namaqualand as basic intrusives and pyroxene granulites (Joubert 1971) around Keimoes as basalts and norites (Poldervaart and Von Backström 1949) and in the Warmbad area as varieties of gabbro (Beukes 1973). The wide range of terms applied to these rocks reflects the problem in finding an origin or them, a difficulty that was also encountered in the present area.

6. *Kum Kum mafic granolite formation*

This formation is restricted to the farm Kum Kum in the extreme west of the area. In hand specimen the rock is melanocratic and practically identical to some versions of the Stolzenfels norites. A crude foliation and a banding defined by thin feldspathic seams are present in the rock and this is the main criterion for distinguishing these mafic granolites from the norites.

In thin section the rock is seen to have essentially the same mineralogy as the norites (Table 10) i.e. ortho- and clinopyroxenes, hornblende, biotite, plagioclase and quartz. The foliation detectable in hand specimen is produced by the alignment of biotite or plagioclase or by the elongation of plagioclase aggregates.

TABLE 10.

Estimated modal analysis of rocks of the Kum Kum formation

Sample No.	Bi	Cp	Op	Hb	Pg	Qz
38	2	6	6		70	8
40	1	8	20		70	2
1460A	5		6	{10}	80	3
1477A	1		30	{10}	60	6
1478	8	5	15		60	8

The mafic granolites are intercalated with rocks of the Jerusalem and Austerlitz formations whereas the contact with the Stolzenfels norite is entirely gradational. In the Warmbad area rocks of this formation have been included with the gabbros (Beukes 1973, pp.268-252) and similar rocks in Namaqualand are described as pyroxene granulites by Joubert (1971, pp.38-42).

7. *Jericho amphibolite formation*

Fairly narrow bands of hornblende gneiss are very common in some areas.

They are referred to here as amphibolites in order to distinguish them from the thicker and more variable sequence of hornblende-bearing rocks forming the Pelgrimsrust formation. There is also a gradation between continuous narrow bands of Jericho amphibolites and larger bodies of hornblende-bearing rocks that have resulted from the retrograde metamorphism of the norites. The amphibolites are composed of hornblende and plagioclase with biotite, epidote, chlorite, perovskite, clino- and orthopyroxene (Table 11). They normally have a dark-grey colour and most of them have a characteristic 'pepper-and-salt' appearance. Contacts with other rock types are sharp and the amphibolites are usually conformable with the foliation in the surrounding gneisses. On the farm Stolzenfels some amphibolites crosscut the foliation and intersect other amphibolite bands in the manner of a dyke and sill complex. Their mineralogy has much in common with the Stolzenfels norites and they probably share a common origin.

TABLE 11.

Estimated modal analysis of rocks of the Jericho formation

Sample No.	Hb	Cp	Op	Bi	Ch	Ga	Po	Pg	Qz
272					{20}		20	40	15
17A	{50}	2						50	3
1922	{20}							70	15
1941	{40}			5				50	1
1761A	{25}			3		8		20	50
1822	{30}	6						50	10
630		15	15					40	15
653					{20}			(50)	10
661B	20	15	5					50	5
784	{40}				{20}			30	10
837A	{40}	5						50	5
919A	{8}	5						(70)	17
1065A					{60}		8	(30)	
1075				10		30		(40)	15
2568	{30}		10	5				50	2
577A	{30}			15				40	10
2341	{60}							30	10
2402	{40}							30	40
2544					{30}				30
2354					{60}		10		30

CHAPTER III

METAMORPHISM

It is beyond doubt that the gneiss sequence under discussion has been subjected to high and even very high grades of metamorphism. It has also been established (c.f. Table 12) that the metamorphic development of the area has been complex. In these circumstances it is essential to establish which criteria may be used with confidence in unravelling the metamorphic history of a 4000 km^2 area in which there is no continuity of outcrop.

A review of the literature shows that several approaches to this problem have been used with little unanimity on basic assumptions. Detailed studies of very small areas have usefully employed the well established relationship between a metamorphic paragenesis and a structural feature such as foliation or lineation. Many authors have extended this technique to larger areas with varying degrees of success (Rast 1963, Joubert 1971, Hopgood and Bowes 1972). It is sometimes assumed for example that a foliation defined by a certain mineral paragenesis in one area is the equivalent of another foliation defined by a similar paragenesis in another area. Alternatively some authors have accepted that a penetrative axial plane foliation may be correlated between separate areas and that any change in mineral parageneses accompanying this foliation reflects changing metamorphic conditions (Skinner et al. 1969, Brown et al. 1969, Pearson 1972).

Using a fabric/mineral paragenesis as a 'datum' over large areas has disadvantages. It is generally accepted that regional metamorphism results from unusually high thermal activity in the crust and many authors such as Zwart (1960) in the Pyrennees and James (1955) in Wisconsin have shown that the extent of these 'thermal domes' may be relatively small ($\sim$10 km) giving rise to rapid changes in metamorphic grade over short distances. From studies in the Alps, Chadwick (1968) has shown that a deformational event may migrate through the lithosphere leaving structures imprinted with a paragenesis in one locality while equivalent structures in other localities have no such associated paragenesis. Also the 'datum method' offers no solution to the problem of distinguishing identical parageneses produced by different metamorphic events widely separated in time.

The acceptance of uniform structural regimes extending over large areas is particularly questionable. Strain is now known to be far more heterogeneous than was previously supposed (c.f. Chapters IV-VI) and it is unlikely that deformational features will be distributed uniformly through large volumes of the crust.

Assumptions about metamorphic and structural correlation methods can only be validly applied in areas of low metamorphic grade. In these cases structural data are often unambiguous and enable primary layering and foliation to be distinguished with relative ease. Mineral parageneses are unequivocal since shallow crustal levels are typically involved. In areas of higher metamorphic grade recrystallisation is pronounced and distinctions between primary and metamorphic layering are lost. At these deeper crustal levels melting and dehydration take place and the fluctuation, by even minor amounts, in the availability of water may profoundly affect these processes (Winkler 1974, p.294). This will also influence the rheological properties of the rock and structural patterns will vary sympathetically (Watson 1973). Structural/metamorphic regimes will not only change considerably with increasing metamorphic grade (Holland and Lambert 1969) but it appears likely that their extent and homogeneity will be reduced as metamorphic grades rise.

The above examples illustrate that the extrapolation of assumptions used in detailed studies to larger areas is not simply a matter of changing scale but raises fundamental questions about the nature of metamorphism since the areas involved are several times larger than the depth of the continental crust.

Some metamorphic petrologists assume that metamorphic mineral parageneses represent the highest grade of metamorphism achieved in the rock concerned (Winkler op. cit. p.17, Thompson 1957). The natural corollaries of this assumption are that:

(i) all regional metamorphism (except shear zones) is prograde (Winkler op. cit. p.17).

(ii) co-existing metamorphic minerals are stable equilibrium assemblages (Winkler op. cit., pp.27-29).

As an approach to field interpretation these assumptions have enjoyed little support. Rast (1965) mentions several regionally metamorphosed areas in Scotland which contain polymetamorphic assemblages (Fig. 6), and Crawford and Oliver (1969) presented an extensive review of areas where an initial high grade metamorphic event was followed by one of lower grade - apparently a common sequence in the Precambrian crust.

Several authors, notably Fyfe (1960) and McKenzie (1965) have criticised the acceptance of textural criteria for indicating stable equilibrium mineral parageneses. Yoder (1952) states that there is no known way of distinguishing between stable and metastable equilibrium conditions. Both Chinner (1961) and Reinhardt (1968) cite examples where, on textural grounds, stable equilibrium parageneses are indicated but in each case one of the minerals involved is known to be a metastable relict phase from a previous metamorphic event.

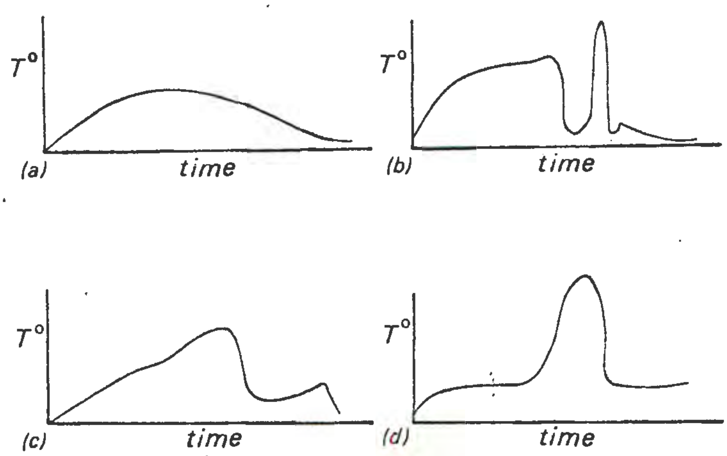


Figure 6. Suggested curves of metamorphic history. (a) theoretical according to Winkler (1974). (b) central Highlands of Scotland. (c) North West Highlands of Scotland. (d) Buchan Dalradians, Scotland. Modified after Rast (1965).

A. APPROACH AND OUTLINE OF METAMORPHIC EVOLUTION

Bearing in mind the limitations of many of the above techniques which are unsuitable for the interpretation of large crustal blocks it was decided to adopt the approach used by Wynne-Edwards (1969) in the mapping of a segment of the Grenville Province in Canada. He emphasised the use of intrusions to distinguish between the metamorphic events and formations preceding them and the metamorphic events that subsequently affected them. On this basis he was able to divide the complex into pre-tectonic, syn-tectonic and post-tectonic categories characterised by rock formations and metamorphic events. Although this approach will not overcome all problems, the writer believes that no other method will so successfully reduce the inconsistencies and subjective interpretations that are inherent in other approaches at this scale of examination.

The writer is fortunate in having within the area of investigations two extensive bodies of clearly intrusive rocks, namely the Beenbreek and Naros granites. Since one of these intrusions (Beenbreek) predates the other, it is proposed that a broad outline of the igneous and metamorphic evolution of the Complex can be established by examining the metamorphic grade of the intruded rocks (especially the xenoliths) and the effects of metamorphism on the younger granite. In using this approach the author applied the following constraints:

(i) Only xenoliths from undeformed granite were examined. This reduces the possibility that later metamorphism has affected the mineral assemblages.

(ii) Xenoliths were not taken from the central part of these bodies. This eliminates the possibility of exotic material confusing the interpretation.

(iii) Xenoliths in contact zones must be clearly derived from the country rock in that locality.

If these conditions are fulfilled it can be accepted that rocks of formations found as xenoliths were in existence before the granite and that the mineral parageneses they contain indicate the metamorphic conditions prevailing prior to the intrusion. This approach relies on the correct correlation of igneous rocks in different parts of the area, which is thought to be an easier task than the spurious identification and correlation of individual foliation surfaces.

The use of this approach revealed a sequence of events outlined in Table 12.

TABLE 12.

Igneous and metamorphic evolution in the Onseepkans area

Igneous Rock	Relationship Observed	Metamorphism	Event
	Foliation and migmatisation of Naros granitoid	M_3	(late)
Naros granitoid			Vellorian (post-Beenbreek)
	Intruded into foliated Beenbreek megacrystic granite and quartzo-feldspathic gneiss (Austerlitz formation)	M_2	(early)
Beenbreek megacrystic granite			
	Intruded into a) garnet-sillimanite gneiss (Jerusalem formation) b) mafic granolites (Kum Kum formation) c) norites (Stolzenfels formaion)	M_1	Kumian (pre-Beenbreek)
Norites			
	intruded into quartzo-feldspathic gneiss? (Austerlitz formation)	M?	

Since areas of low strain preserving undeformed marginal zones of syntectonic granites are not common the collection of xenoliths that satisfy the above constraints is difficult. Therefore, for the vast majority of the area (i.e. non-xenolith environments), it is not possible to describe the rocks as being composed of pre- and post-Beenbreek assemblages. It will be assumed, therefore, that parageneses over the area as a whole result from the combined effects of the different events. In some instances these 'compound' parageneses may be further evaluated but the general absence of more than one fabric in any rock and the paucity of significant mineral relationships described from other areas (Zwart 1960, Johnson 1963) normally make this impossible.

The following discussion on metamorphism is hindered by the lack of quantitative data. No whole rock chemical analyses are available from the study area and only in one instance was a mineral composition identified. However, this thesis represents a field oriented study and such investigations lie outside the frame of reference. The reader must therefore judge the validity of the conclusions with these limitations in mind.

B. PRE-NORITE METAMORPHIC EVENT

One exposure was found on the farm Eendorn where norite apparently intruded quartzo-feldspathic gneisses of the Austerlitz formation. The foliation in the gneiss appeared to terminate against the norite but no banding or layering was crosscut. This relationship is not convincing engough to propose a pre-norite metamorphic event since it is open to other interpretations (c.f. Chapter IV). Joubert (1971, p.61), found similar equivocal contacts in Namaqualand although Clifford et al. (1975) cite reliable evidence for the intrusion of similar noritoids into metamorphic rocks in the O'okiep area. In the Onseepkans area it is therefore proposed that the development of the complex commenced with norites and the associated regional metamorphic event.

C. PRE-BEENBREEK METAMORPHIC EVENT (KUMIAN)

Of the several rock types found as xenoliths in the Beenbreek granite the most significant are the Stolzenfels norites and the Kum Kum mafic granolites. Four major mafic bodies of the Stolzenfels group are present in the area and, for convenience, these are numbered SI to SIV. Body SI occurs on the farms Kum Kum and Eendorn, SII on Keimasmond and Orangefall, SIII on Ondermatjie and Jerusalem and SIV on Stolzenfels and Jericho. Many smaller bodies are present, especially on the farms Vaaldoorn and Beenbreek. There is conclusive field evidence that SI, SII and minor norites on the farm Beenbreek predate the Beenbreek granite. SI is associated with a mineralogically indistinguishable rock, the Kum Kum mafic granolite, which also occurs as xenoliths in the Beenbreek granite. The only other rocks found as xenoliths in the granite are gneisses belonging to the Jerusalem formation.

1. Stolzenfels norite

Rocks of this formation were described in section IIB5. Most thin sections of these rocks show two pyroxenes. The orthopyroxene is hypersthene often weakly pleochroic from pale green to pale pink. Schiller structure, defined by thin plate-like inclusions of opaque minerals in the cleavage is common. The crystal habit varies widely from subhedral grains less than 1 mm in size to large, very irregular and sometimes fishnet crystals. In one thin section the hypersthene was in optical continuity.

The clinopyroxene is augite. Weak pleochroism is sometimes present and the crystal habit has a variation similar to hypersthene. Twinning is rare and undulose extinction was seen in two thin sections from SI. Both clino- and orthopyroxene appear to be stably co-existing phases and are not derived from any prograde metamorphic reactions. A typical ophitic texture was seen in one thin section (DT1512).

Plagioclase varies between andesine and labradorite. Albite and pericline twins are invariably present. Only a few crystals are anti-perthitic. A weak alignment of plagioclase crystals can sometimes be seen, and this may be accompanied by deformed twin lamellae. Very little microcline was identified.

Hornblende with a dark-brown colour apparently co-exists stably with pyroxene in two thin sections. The colour appears to be characteristic of the temperature of formation (Miyashiro 1973, pp.254-257). Brown hornblende forms at high temperatures, green hornblende at intermediate temperatures (amphibolite facies) and blue-green hornblende at low temperatures. A green hornblende is found as reaction rims around pyroxene in many thin sections. Biotite and quartz are common accessory minerals. Over 20% quartz occurs in some samples of SIII.

Igneous rocks containing hypersthene are now widely recognised as belonging to the charnockitic rock group (Howie, 1964, Tobi 1971). They characteristically occur in Precambrian terrains, possess a distinctive dark-grey, greasy appearance (Howie 1967, Oliver and Schultz 1968) and are usually associated with high-grade metamorphic rocks. The compositions of the Stolzenfels norites are plotted in Fig. 7 taken from Tobi (op. cit.) and modified in discussion with Torske (1972). This diagram conforms closely with the Streckeisen classification for normal igneous rocks. The majority of specimens plot as norites with the exception of some samples from SIII which extend into the enderbite field.

One hand specimen from the Stolzenfels norite (SI) and one from the farm Vaaldoorn were found to be olivine gabbro but no actual contacts with the norites were found. They are black, coarse-grained, and composed of olivine, clino-and orthopyroxene, plagioclase and biotite. The sample from Vaaldoorn had an ophitic texture. The sample from SI displayed well developed coronas around olivine crystals in contact with plagioclase. The coronas are composed of an inner core of olivine surrounded by a rim of orthopyroxene and enclosed by a symplectic intergrowth of clinopyroxene and green spinel (Plate 11).

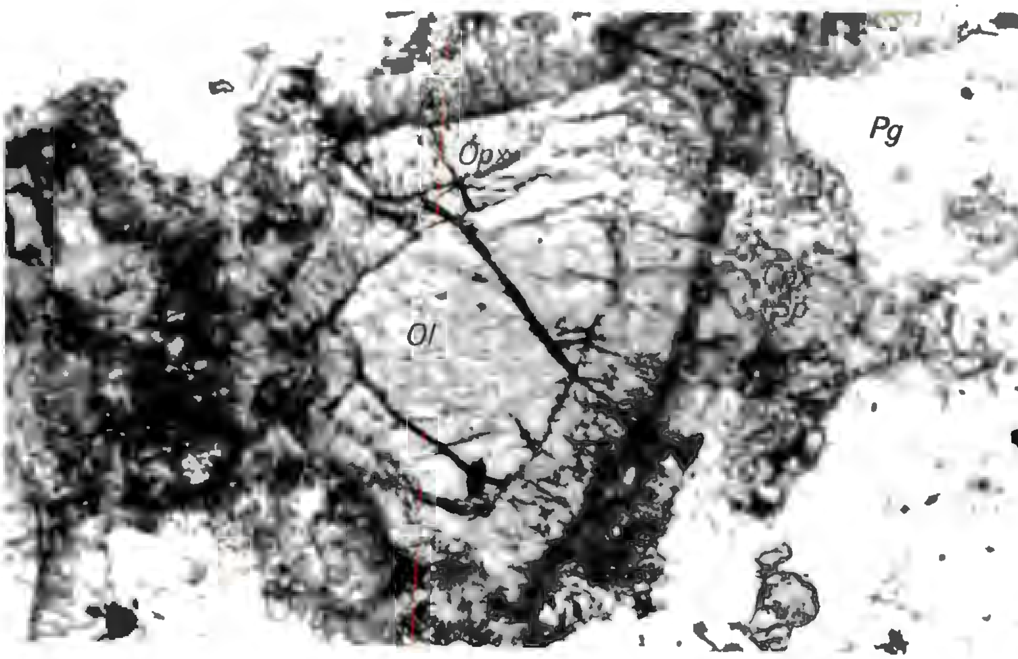


Plate 11. Photomicrograph of olivine gabbro from SI norite body. Reaction rims around olivine are composed of an inner rim of orthopyroxene and an outer rim of clinopyroxene and green spinel. Keimas Farm.

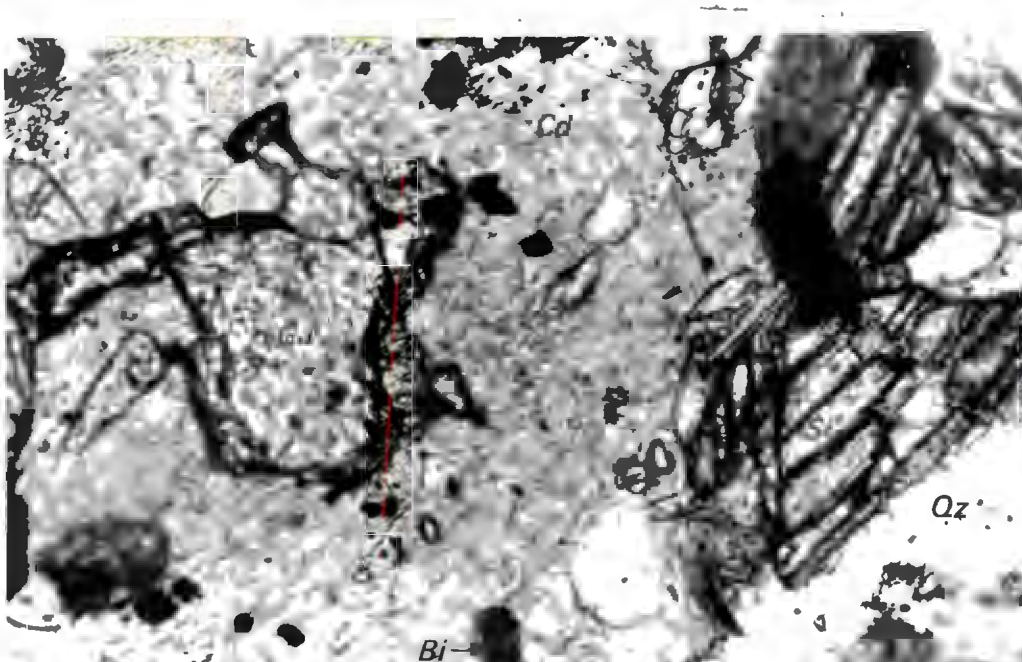


Plate 12a. Photomicrograph of Jerusalem gneiss showing cordierite reaction rims around garnet. X40. Kum Kum Farm.

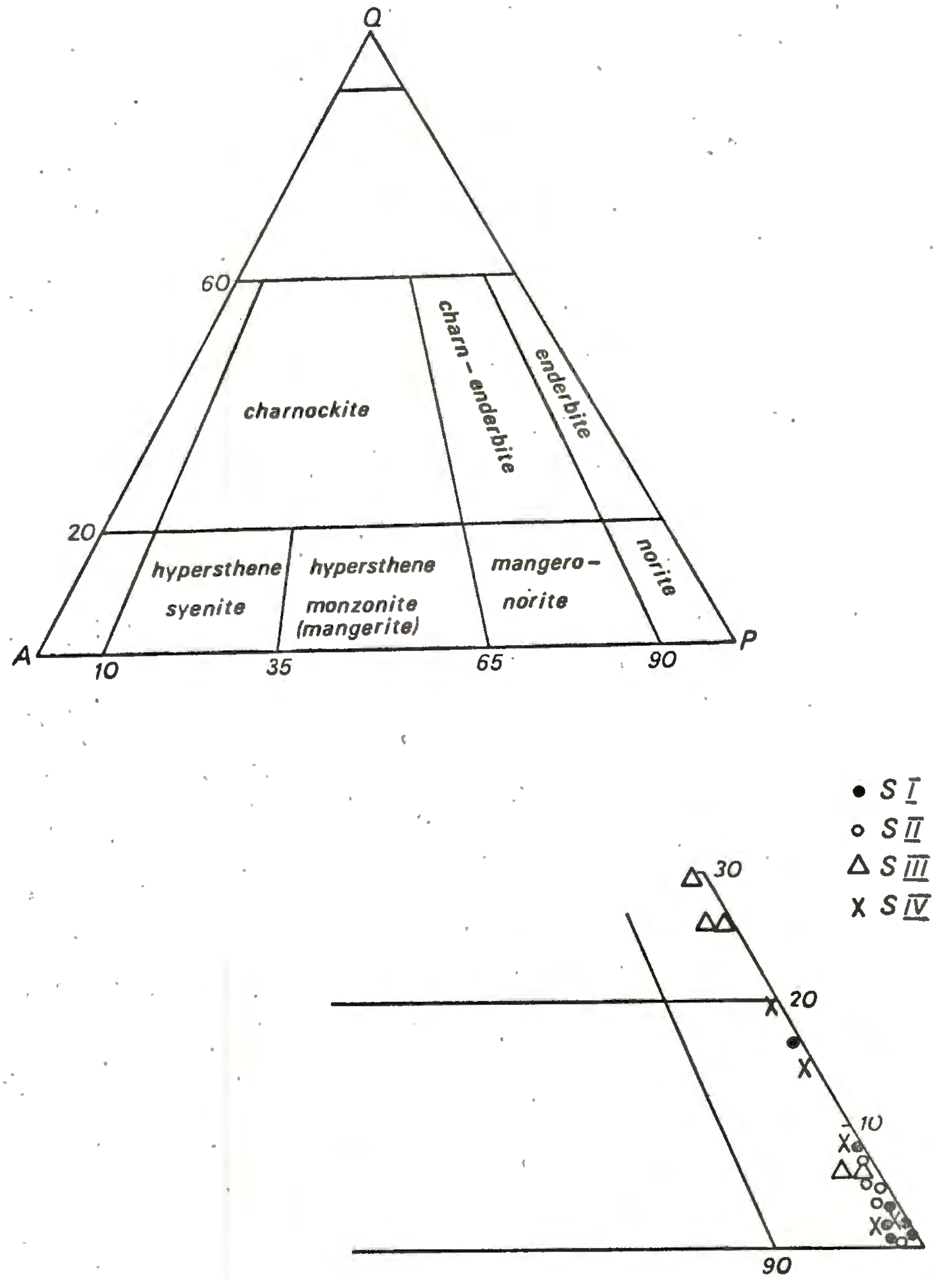


Figure 7. Composition of the Stolzenfels norites and enderbites plotted on a classification diagram taken from Tobi (1971, 1972).

Whitney and McLelland (1975) have described identical coronas in olivine gabbros from the southern Adirondaks and they ascribe this phenomenon to prograde metamorphism (i.e. *ol* + *pg* ⇌ *cp* + *op* + *sp*) in otherwise typical igneous rocks with ophitic textures.

2. *Kum Kum mafic granolite*

This formation of granulite-grade metamorphic rocks is restricted to the extreme west of the area on the farm Kum Kum. They are greasy looking dark-grey fine-grained rocks that are virtually identical in hand specimen to rocks from the norite. They are distinguished by a granoblastic elongate texture or sometimes a crude banding defined by feldspathic aggregates. Mineralogically they are identical to the norites (Table 10) and typical parageneses are:-

plagioclase - orthopyroxene - biotite - quartz±hornblende

plagioclase - orthopyroxene - clinopyroxene - biotite - quartz±hornblende

They are shown in an ACF diagram in Fig. 8. The term granolite is preferred for these hypersthene-bearing rocks rather than granulite, following a proposal by Winkler (1974, p.247). Previous attempts to reach international agreement on the definition of the term granulite proved unsatisfactory (Behr et al. 1971, Mehnert 1972) and the term is used here for a high grade of metamorphism of which granolites are the diagnostic rock type.

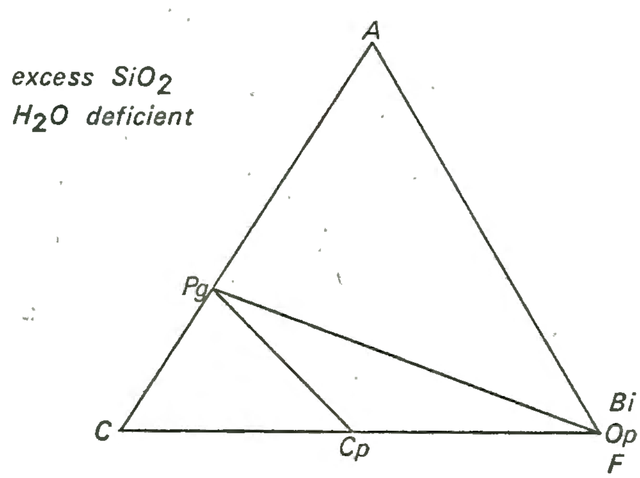


Figure 8. ACF diagram for rocks of the Kum Kum formation showing Kumian parageneses.

Under the microscope the granolite foliation can be related to the vague alignment of biotite and plagioclase crystals which are often characterised by deformation twins. Occasionally it is defined by the elongation of plagioclase aggregates. The two pyroxenes commonly show no sign of retrogression indicating that the fabric was impressed upon the rock under P-T conditions suitable for the stable co-existence of these minerals. Since the foliation direction is shared by the intercalated Jerusalem gneisses it is unlikely that it originated as a primary magmatic feature.

Interpretation

The origin of charnockitic rocks is open to question. Some authors have maintained that they are igneous (Subramaniam 1956, Philpotts 1966) while others have suggested their origin may be solely metamorphic (Cooray 1969). Sen and Ray (1971) and De Waard (1969, 1973) proposed an igneous origin modified by granulite grade metamorphism.

At Onseepkans the association of olivine gabbro and norite and the presence of ophitic textures strongly support the hypothesis that these rocks were originally igneous. However, the existence of mafic granolites grading into the norites and identical in every respect to the norite except for a tectonic fabric points to a granulite-grade event taking place either contemporaneously with or following the intrusion.

From studies on metabasic rocks in the Adirondack Mountains, De Waard (1964, 1965a) has suggested that the limiting reactions for the granolite assemblage *cp - op - pg - qz* are:

$$op + cp + pg \rightleftharpoons hb + qz \text{ (amphibolite facies)}$$

$$op + pg \rightleftharpoons cp + al + qz \text{ (high-pressure granulite facies)}$$

The absence of almandine and primary hornblende (i.e. an hydrated phase) would confine these mafic granolites to a low or intermediate pressure subzone (De Waard 1965b) of granulite-grade metamorphism.

The reaction $ol + pg \rightleftharpoons cp + op$ implied by the coronas within the olivine gabbro sample from SI and investigated experimentally by Green and Ringwood (1967, p.818) has been cited as marking the transition from low to intermediate pressure granulite grade. The same authors have also related the paragenesis $op + cp + pg + qz$ to an intermediate-pressure granulite subzone.

3. *Jerusalem cordierite-sillimanite-garnet gneiss formation*

Gneisses of this formation are found as xenoliths in the Beenbreek granite at the junction of the farms Kum Kum and Keimas. They are also intercalated with the mafic granolites on Kum Kum where they share a very similar dark-grey colour and greasy appearance. Typical parageneses are:

quartz + sillimanite + garnet + cordierite + biotite ± microcline ± plagioclase

quartz + sillimanite + garnet + biotite ± microcline ± plagioclase

They are illustrated in the AFM diagram in Fig. 9.

Following the observations of Wynne-Edwards and Hay (1963) it is now well established from field and experimental data that the pair cordierite-almandine is virtually restricted to high-grade metamorphic rocks. The co-existence of these minerals is controlled by several possible reactions which are influenced by the MgO + FeO/Al_2O_3 and MgO/(MgO + FeO) ratios of the rock, and the Ca and Mn content of the garnets. However, one point that has emerged from the studies of Hensen and Green (1971, 1972, 1973) and Currie (1971, 1974) is that the field of co-existence is strongly pressure-dependent.

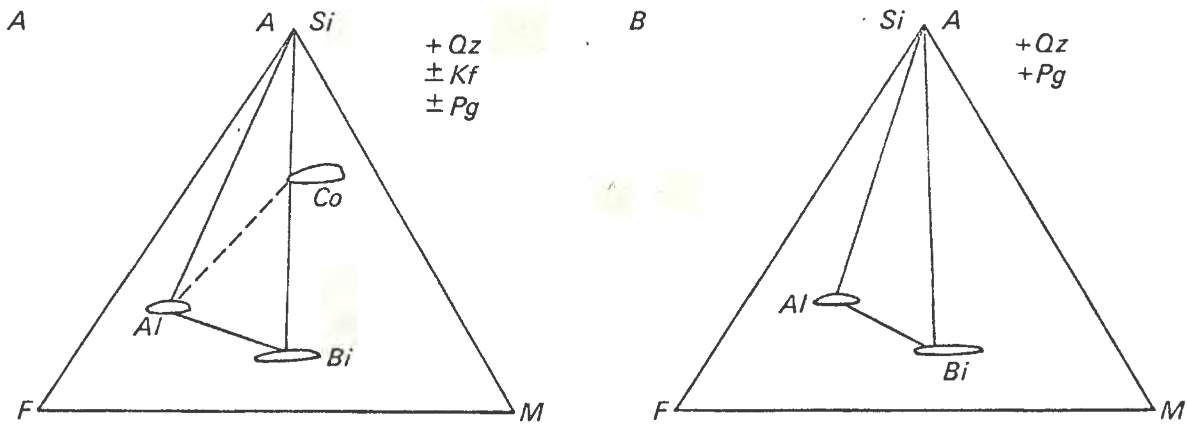


Figure 9. AFM diagrams for rocks of the Jerusalem formation showing Kumian parageneses. Modified after Reinhardt (1968).

Under the microscope typical minerals include coarse-grained sillimanite (up to 1 mm in diameter) large, sometimes poikilitic garnets (up to 2 mm), fox-red biotite, pinitised xenoblastic cordierite and non-perthitic plagioclase and microcline. The origin of cordierite in these rocks is an interesting problem since the mineral appears to be derived from the reaction of several other phases. According to Reinhardt (1968) the assemblage *al + cd + bi + si* is unstable and a reaction between some of these constituents is to be expected. In almost all thin sections of Jerusalem gneiss the cordierite appears in very minor amounts and invariably form incomplete reaction rims around the garnet. The only cordierite-forming reactions involving garnet are the divariant reactions:

$al + si + qz \rightleftharpoons cd$ (Hensen and Green 1971, p.325)

$al + kf + H_2O \rightleftharpoons cd + bi$ (Currie 1971, p.218)

Although microcline is sometimes present it is rarely seen in contact with cordierite and therefore the first of the above reactions would appear more applicable. Plate 12a illustrates a typical paragenesis involving cordierite. In addition to garnet, cordierite encloses biotite, ore, quartz and sillimanite. Sometimes small grains of cordierite are found without a garnet core (Plate 12b) and since no other phases except biotite, quartz and sillimanite are present, the absence of garnet may be due to the reaction $al + si + qz \rightleftharpoons cd$ having gone to completion. Coronas of garnet on cordierite are best explained by referring to the diagrams of Currie (1971, p.224) and Hensen and Green (1971, p.329), reproduced in Fig. 10. It can be seen from these diagrams that increasing pressure changes a cordierite-bearing paragenesis to cordierite + almandine and eventually to an almandine-bearing assem-

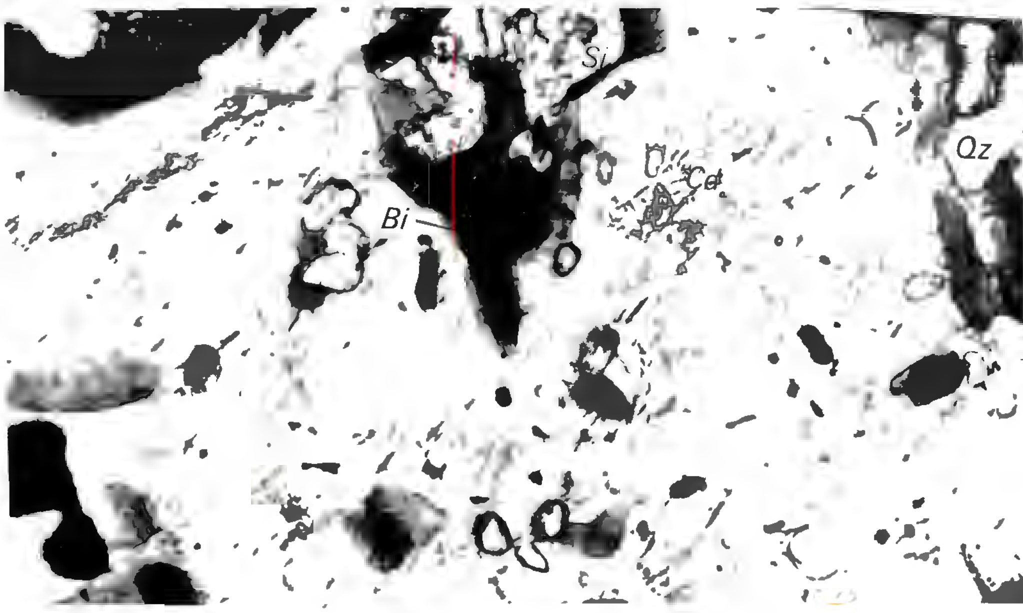


Plate 12b. Photomicrograph of Jerusalem gneiss showing cordierite enclosing biotite, sillimanite and quartz. X40. Kum Kum Farm.

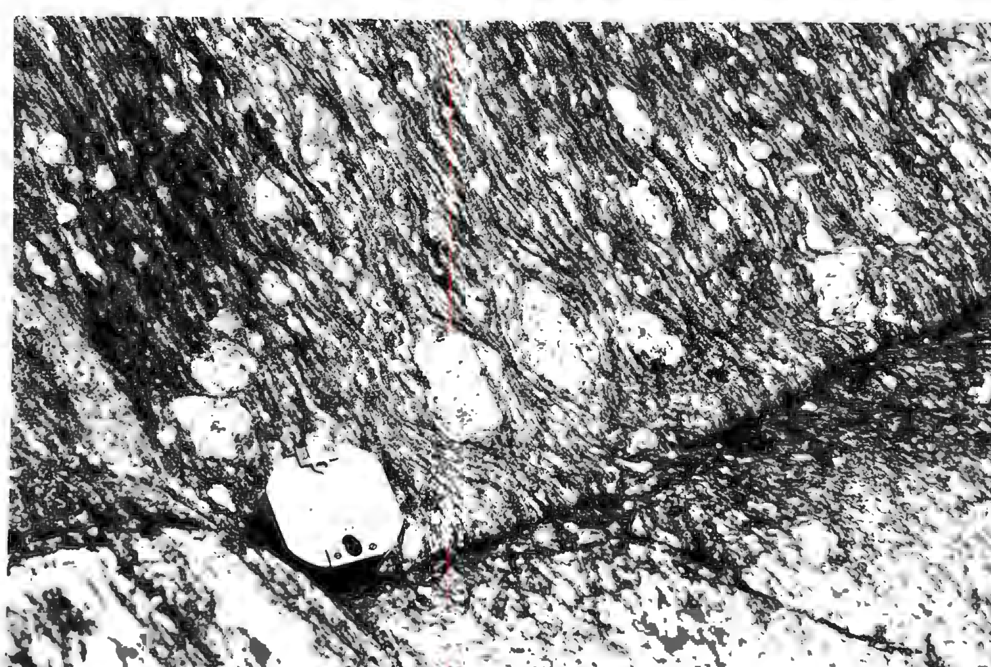


Plate 13a. Deformation of Beenbreek megacrystic granite producing a foliated granite. Beenbreek Farm.

blage by the reactions quoted above. Cordierite rimming garnet may therefore be produced by the reverse of this process under conditions of diminishing pressure, i.e. the assemblage is moving from the higher pressure almandine field into the lower pressure field of cordierite-almandine co-existence, as shown schematically by the dashed arrows in Fig. 10

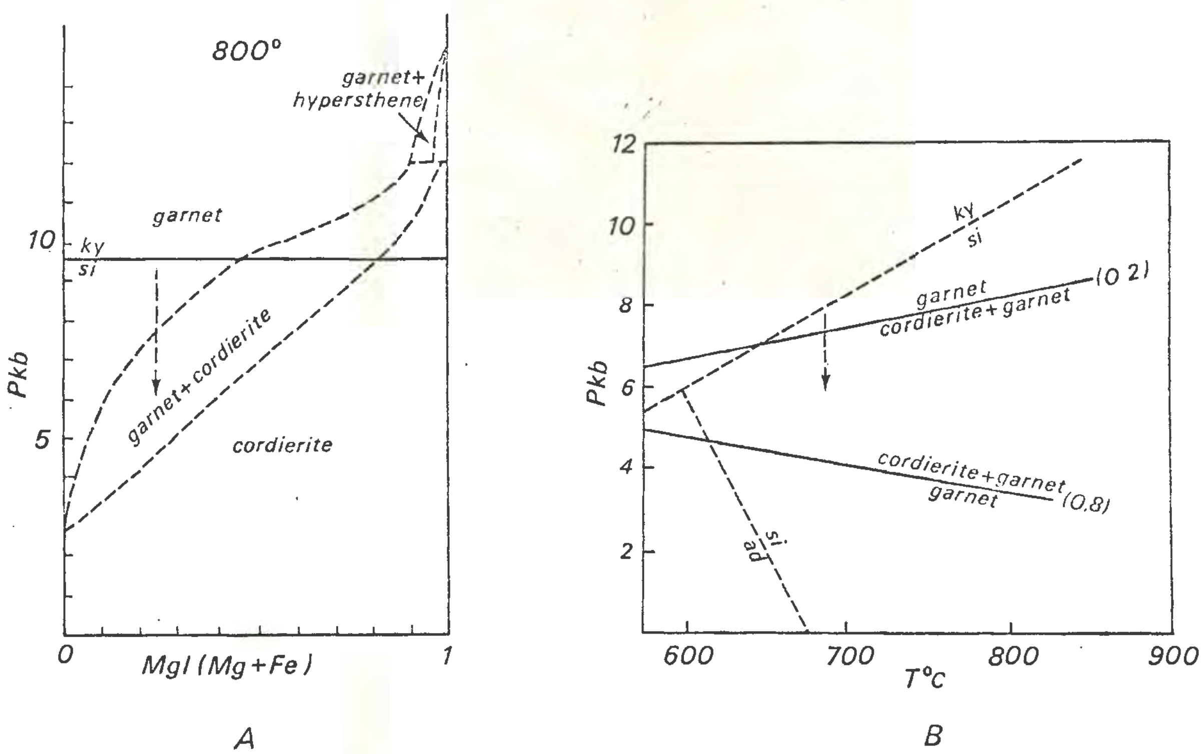


Figure 10. Stability fields of garnet and cordierite. A. Hensen and Green (1971) at $800^{\circ}C$. B. Currie (1971) for Fe/(MgO+FeO) ratios between 0,8 and 0,2. Dashed arrows in both diagrams show probable mode of formation of cordierite reaction rims around garnet.

A thin section from a xenolith in Beenbreek granite on the farm Keimas showed garnet but no trace of cordierite (paragenesis B, Fig. 9). Considering the above relationship in the coronites this xenolith furnishes good evidence that the original Kumian assemblages did not contain cordierite. The reaction rims were therefore formed during a later metamorphic event which may have also produced the mantling of pyroxene by hornblende in the mafic rocks. Rims of cordierite around garnet have been reported by Hess (1971) from Massachusetts and Frejvald (1974) from Bohemia. Frejvald ascribed this relationship to later low-pressure re-metamorphism of granulite-grade gneisses.

Inclusions of Jerusalem gneisses in the Beenbreek granite are very common at this locality on Keimas farm and immediately adjacent to the contact are extensive exposures of the Jerusalem formation. Not a single xenolith shows evidence of migmatisation while the adjacent country rock is well migmatised with the development of numerous neosome bands. The foliation in the migmatite is also penetrative through most of the granite. This shows quite conclusively that the migmatisation, at least in this area, post-dated the Beenbreek granite and was not developed during the Kumian metamorphic event.

P-T Conditions of Kumian Metamorphism

Experimental data for several of the reactions and parageneses visible in thin section are available in the literature and can therefore be used to define the approximate P-T conditions for Kumian metamorphism. In resumé these are:

(i) stable co-existence of ortho- and clinopyroxene in granolite (i.e. no garnet present).

(ii) almandine-bearing 'pelitic' assemblages (with possible inclusion of limited cordierite reaction rims).

(iii) olivine + plagioclase $\leftrightarrows$ clinopyroxene + orthopyroxene.

(iv) sillimanite present as the stable $Al_2\ SiO_5$ polymorph.

Hewins (1975) investigated the pair ortho- and clinopyroxene from many occurrences using the geothermometer of Wood and Banno (1973), he concluded that the overwhelming majority of these fall in the temperature field 780^{o}-860^{o}C.

The field of co-existence of garnet and cordierite is wedge-shaped on the P-T diagram (Currie 1971). The upper pressure limit of this field for rocks of average Fe/Mg ratios varies between 7-9 kb at temperatures between 600^{o} and 900^{o}C. This approximate limit was verified by Hensen and Green (1974, p.160) who analysed co-existing cordierite and garnet from several areas.

The usefulness of the olivine + plagioclase reaction is diminished by incompatible experimental results. Problems are traceable to initial compositions investigated and to the extrapolation of these results at lower P-T conditions. Green and Ringwood (op. cit.) proposed an intermediate-pressure granolite subzone (defined by the breakdown of olivine) whose bounding limits are approximately parallel. Kushiro and Yoder (1966) have suggested that this zone is wedge shaped, the two-pyroxene field derived from the breakdown of olivine closing at about 6 kb at 700^{o}C. This has been confirmed by Green and Hibberson (1970) who believe that the size of this field may be even smaller when considering other possible compositional factors. Whitney and McLelland (1975) have put the P-T conditions of the Adirondack coronites at 8 kb and 800^{o}C, a conservative figure when taking these conflicting data into account.

Experimental work by Althaus (1967, 1969) and Richardson et al. (1969) shown that the presence of sillimanite is compatible with the P-T conditions indicated by other assemblages. The transition to kyanite takes place at

6-13 kb between 600° and 900°C (Althaus 1967).

This information is plotted in Fig. 11 which indicates P-T conditions for the Kumian of 8-9 kb at 800-860°C.

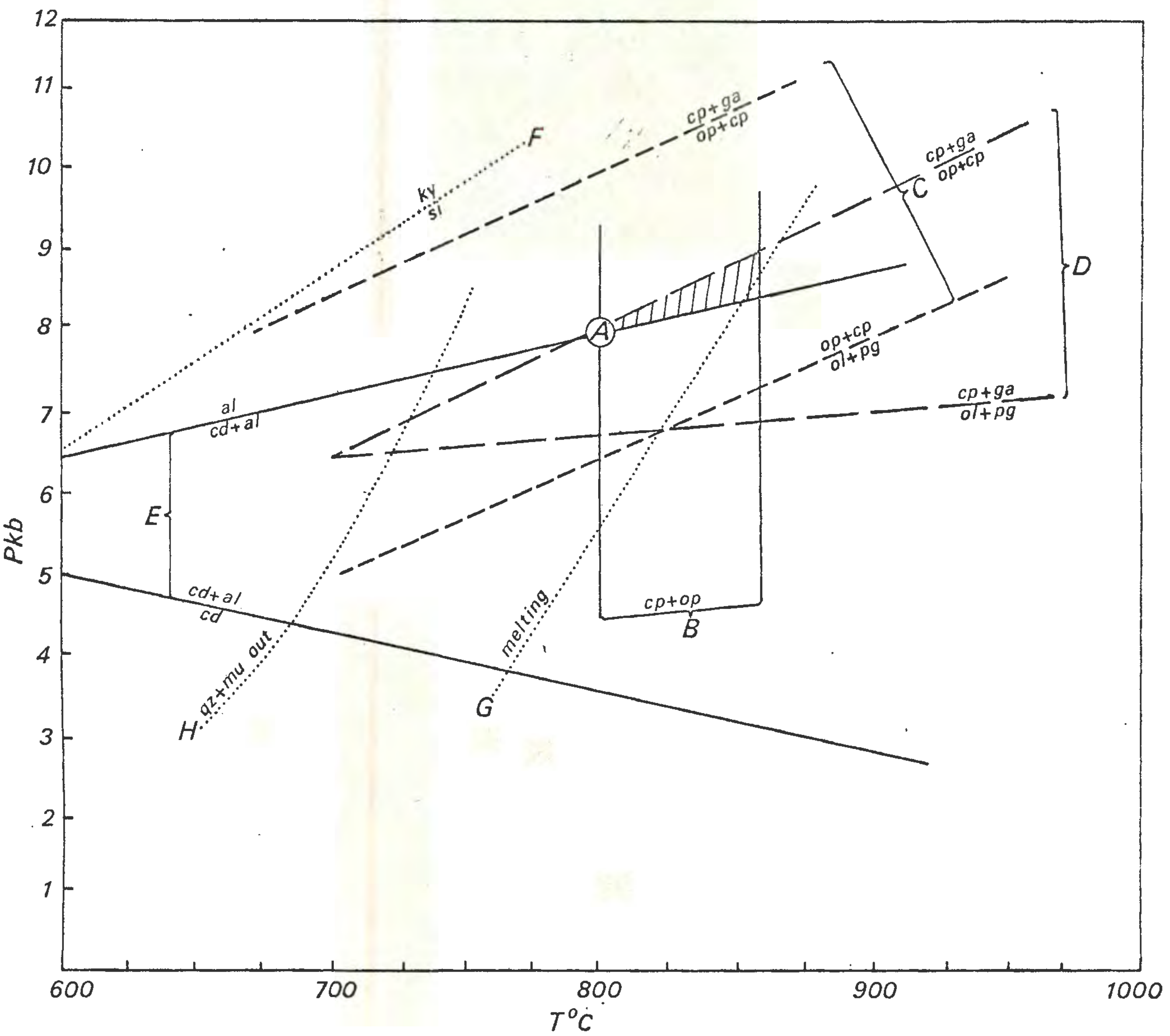


Figure 11. P-T field for Kumian metamorphism (hatched). For reactions see text. A. Whitney and McLelland 1975. B. Hewins 1975. C. Green and Rignwood 1967. D. Kushiro and Yoder 1966, Green and Hibberson 1970. E. Currie 1971. F. Althaus 1967. G. Althaus 1968, PH_2O=1kb for beginning of melting. H. Althaus 1970.

The marked absence of migmatites in these rocks implies extremely anhydrous conditions ($PH_2O<<P$ Total) and raises certain problems, especially with gneisses of pelitic composition. It has been suggested by Winkler (1974, p.262) that these conditions only prevail at deeper crustal levels where water was slowly lost during a long metamorphic history. Since the evidence indicates that the Kumian was the first metamorphic event, this suggestion would not seem appropriate. Touret (1971) has proposed that the same anhydrous conditions could be simulated if the fluid phase contained appreciable quantities of CO_2. He has suggested that the CO_2 may originate through degassing of the mantle. An attempt was made to gain a qualitative estimate of the fluid phase composition by examining suitable inclusions in quartz grains which contain both a fluid and a gas phase. These inclusions are carefully watched while the thin section is heated with a hair drier (Touret, pers. comm.). If the bubble, i.e. gas phase, in the inclusion shows signs of enlarging during the heating it indicates the presence of CO_2 and the degree of enlargement of the bubble is a guide to the relative amounts of CO_2 and H_2O present. The inclusions tested by the writer showed complete homogenisation when heated to approximately 35^oC which can only happen if both fluid and gas are nearly pure CO_2. If the temperature of the thin section is known the composition of the inclusion can be estimated with a high degree of accuracy. Touret (1974) presents tables showing the amount of homogenisation for different compositions at known temperatures.

D. PRE-NAROS METAMORPHIC EVENT (EARLY VELLOORIAN)

In exposures along the Ham River undeformed Naros granitoid can clearly be seen to intrude foliated Beenbreek megacrystic granite (Plate 9) and quartzofeldspathic gneisses. The tectonism following the Kumian and predating the Naros granitoid will be referred to as M_2 - the first of two metamorphic episodes constituting the Velloorian. Only the Beenbreek granite has been found as xenoliths in the Naros granitoid and, recalling the constraints imposed in the introduction (section IIIA) nothing can be said about the effects of M_2 on the Onseepkans sequence. Two topics will be discussed in this section namely the intrusion and deformation of the Beenbreek granite and the formation of the Grass River augen gneiss.

There was a marked change in metamorphic conditions following the Kumian which witnessed the intrusion of an hydrated rock (the Beenbreek granite) and its subsequent deformation. No evidence of migmatisation was found in deformed granite xenoliths which suggests physico-chemical conditions during the deformation were not suitable for anatexis. On the basis of the limited number of xenoliths seen in only one locality this conclusion may, however, not be applicable to the whole area.

(i) *Beenbreek granite*

The megacrysts in the Beenbreek granite consist of either microcline or plagioclase (andesine) or aggregated of these minerals with quartz. Some are weakly poikilitic, containing sub-hedral plagioclase crystals. No inclusions or garnet or sillimanite were found and no megacrysts were seen enclosing a tectonic fabric. The origin of megacrysts in "porphyritic" granites *vis à vis* porphyroblasts in surrounding rocks has been debated for several decades but most authors now appear to accpet blastesis as the origin for both (Mehnert 1968, p.289-295; Marmo 1971, pp. 42-48; Didier 1973, pp.208-215). Beukes (1973, p.212) invoked potassium metasomatism for the origin of microcline megacrysts in his Eendoorn granite where, apparently, plagioclase megacrysts do not exist.

Biotite and garnet are the most common mafic minerals. Although some authors have suggested that garnet in granite is a primary mineral (Hall 1965; Green and Ringwood 1968) it is generally accepted (e.g. Mehnert 1968) that, if the country rocks are rich in garnet, as in this case, its presence is most probably due to assimilation. The occurrence of sillimanite in some thin sections also suggests that the granite was derived by anatexis although it has been suggested that this mineral may be primary (Kennan 1972). Cordierite was found in two thin sections occurring in exactly the same manner as in Kumian assemblages, i.e. as thin, incomplete selvages around garnet. The mineral was only found in thin sections containing both garnet and sillimanite showing that the reaction $al + sil + qz \leftrightarrows cd$ (c.f. section IIIC3) is also relevant in this case. The absence of cordierite as a discrete phase in the granite is additional evidence that it was not formed in the Onseepkans sequence during the Kumian event.

The origin of aplite in the Beenbreek granite raises a further problem since the aplite contains no hydrous minerals. Leake (1974) has described a similar megacrystic granite/garnetiferous aplite association in the Galway Granite, Ireland, and has shown that the aplite is derived from a water-poor residual liquid. If this model is applicable to the Beenbreek granite it indicates that more anhydrous conditions were again present at the end of the intrusive period and this may explain why the subsequent deformation of the granite was not accompanied by anatexis. Fluctuation in the availability of water has been shown by Starmer (1972) to produce periods of granitisation during otherwise anhydrous, granulite-grade metamorphic events.

The foliation of the Beenbreek granite results in a complete gradation from augen gneiss to garnet-biotite gneiss (Plate 13). A similar gradation exists between aplite-bearing granite and augen gneiss containing bands of quartzo-feldspathic rock (Plate 14). On the farm Keimas undeformed megacrystic granite carrying xenoliths of over 1 m in diameter is transformed within a distance of 5 m into a foliated augen gneiss in which the drawn out and flattened xenoliths are approximately 15 cm thick. Surprisingly, the megacrysts in such cases show very little evidence of deformation. The strain effects are usually limited to bending of twin lamellae and a weak mortar structure developed round the edges of the augen. At one stage the writer felt that the general absence of deformational features suggested an alternative origin for the Beenbreek granite (Toogood 1974). In this model the augen gneisses were seen as developing by blastesis in suitable host rocks and the granite was seen as the localised end product of this process giving rise to restricted intrusive phenomena. This hypothesis is now rejected on the following grounds:

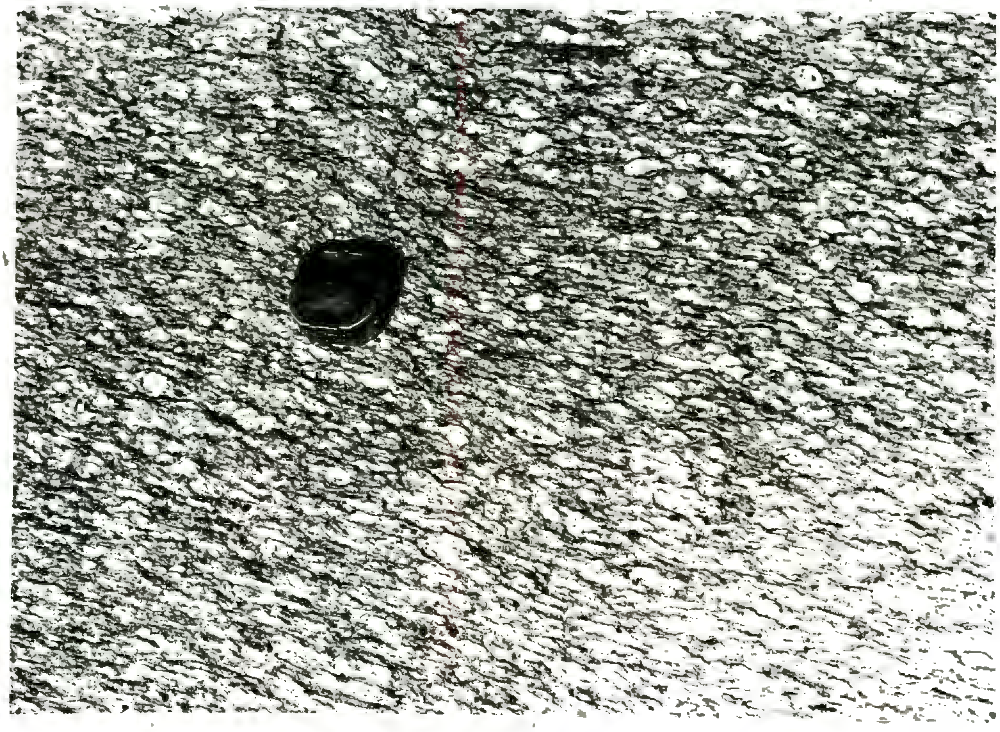


Plate 13b. Deformation of Beenbreek megacrystic granite producing an augen gneiss. Beenbreek Farm.

Plate 13c. Deformation of Beenbreek megacrystic granite producing a garnet-biotite gneiss. Ondermatje Farm.

(i) very few of the megacrysts are poikilitic and none shows a tectonic fabric.

(ii) no xenoliths of augen gneiss (i.e. an intermediate stage) have been found in the granite.

(iii) contacts between granite and augen gneiss are gradational and not intrusive.

(iv) aplite intruding granite probably not produced by the mobilisation of intercalated augen gneisses and garnetiferous quartzo-feldspathic gneiss.

(v) cases exist where granite with xenoliths grades into augen gneiss containing narrow lense-like horizons of other rock types. This could not have been produced by mobilisation of the augen gneiss.

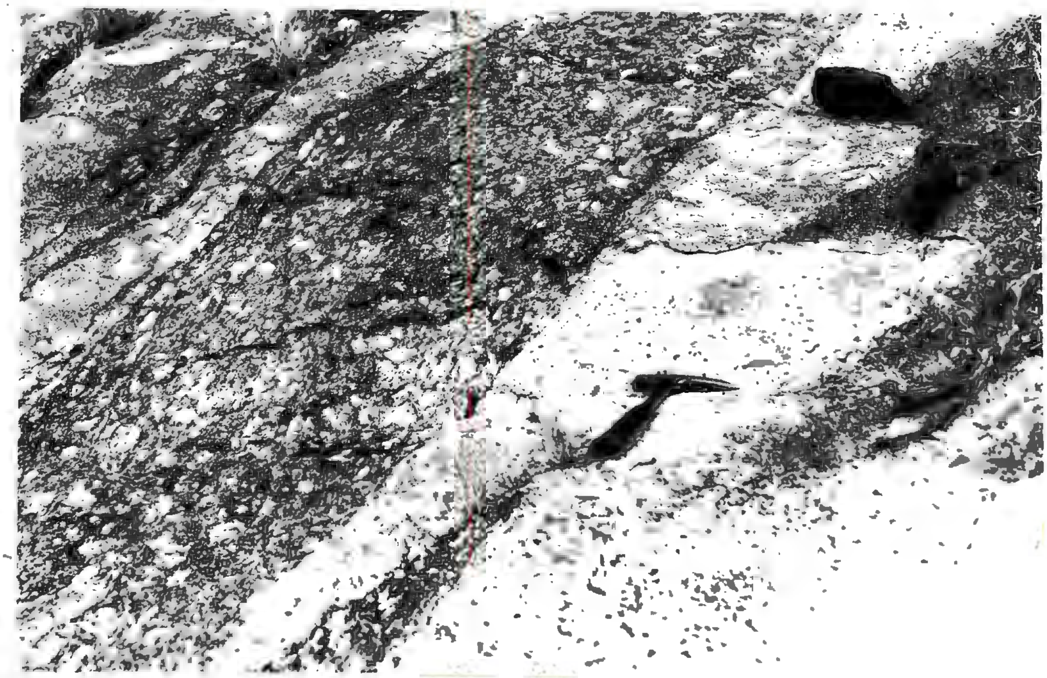
On these grounds it was concluded that the augen gneiss results from the penetrative deformation of a previously isotropic megacrystic granite.

2. *Grass River augen gneiss*

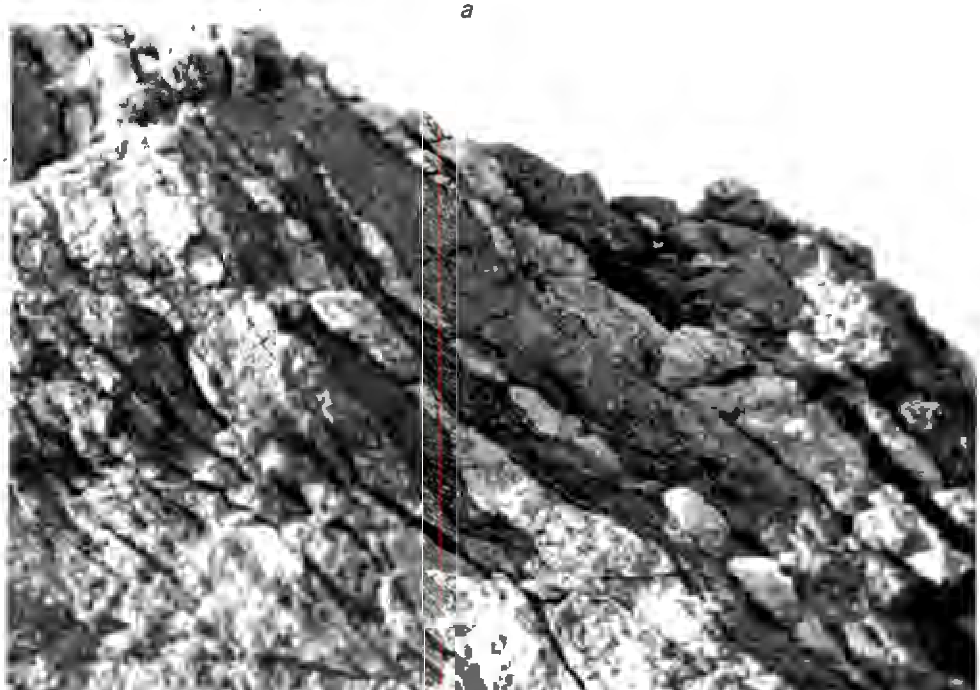
Little can be added to the lithological description of this formation as given in section IIB2 since it has not been possible to determine whether these rocks are largely composed of blastites or of deformed megacrystic granite. In some instances compositional variations fall outside the normal range for Beenbreek granite and therefore an origin through blastesis is more tenable. Such examples include augen gneisses in marginal parts of the Stolzenfels formation (especially the eastern contact of SII) and, also, the augen gneisses on the farm Blyderverwacht which are considerably more leucocratic than the Beenbreek formation. Unfortunately augen are not sufficiently distinctive to be of use, although they are potentially very valuable strain markers (Wilson 1972).

Theoretically megacrysts can be of four main types. In Fig. 14a, a porphyritic fabric is developed in an igneous rock with unoriented crystals set in an isotropic matrix. Apart from some undeformed parts of the Beenbreek granite where megacrysts could be interpreted as phenocrysts, this fabria is unknown from the present study area. Fig. 14c. illustrates a classical, unambiguous case of a porphyroblast growing in, and replacing, a previous tectonic fabric. Such porphyroblasts are not seen in the Onseepkans area. The augen normally encountered are illustrated in Figs. 14c and 14d, by far the greater majority is represented by Fig. 14d. In these fabrics the augen lie within and do not cut the foliation which curves round them. Occasionally (Fig. 14c) the foliation gives the appearance of abutting against the megacrysts and even continuing some distance into them. These features allow various interpretations.

It has long been maintained that a growing porphyroblast has the ability to displace the surrounding matrix. This view was expressed by



a



b

Plate 14. Deformation of aplite-bearing Beenbreek megacrystic granite producing foliated granite and aplite (a) and ultimately augen gneiss with boudinaged quartzo-feldspathic horizons (b). Compare with underformed relationship in Plate 4. Plate (a) Ondermatje Farm, Plate (b) Jerusalem Farm.

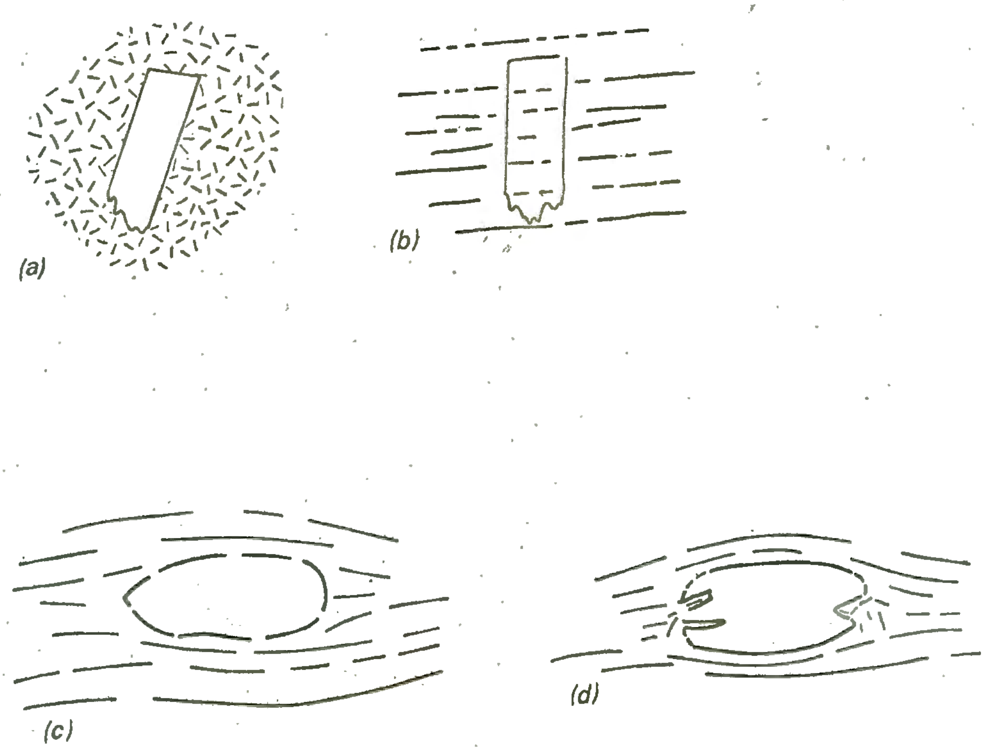


Figure 14. Various types of fabric containing megacryst. (a) phenocryst in igneous rock. (b) porphyroblast replacing a tectonic fabric. (c) augen of uncertain origin representing the majority of megacrysts found in the Grass River augen gneiss. (d) augen of uncertain origin representing a minority of megacrysts found in the Grass River augen gneiss.

Harker (1932) and Ramberg (1947) and apparent proof of this was recently furnished by Saggerson (1974). However, as pointed out by Spry (1969) and Spry and Misch in discussion (1972) "no rock which has suffered deformation concurrently with, or subsequent to, metamorphism can be used to prove displacement (by porphyroblast) convincingly". It can be readily shown that the Grass River augen gneiss have undergone two periods of penetrative foliation. Therefore, any information gained from the present textural features of the megacrysts is of little value in deciding whether the rock originated as an intrusive granite or as a blastite.

It may be proposed here that post-megacryst deformation must inevitably be reflected in the strain features within the megacrysts themselves. This was not the case in the Beenbreek granite where augen are known to represent deformed megacrysts and the situation in the Grass River formation is virtually identical. Very little published information is available on deformational textures in feldspars (White 1975) and this makes it very difficult to hypothesize on the reasons for their absence in the augen gneiss megacrysts. It is tentatively suggested that the penetrative foliation, or refoliation, of a rock containing megacrysts in a low strain rate situation will not necessarily produce these deformation textures. However under high strain rate conditions mortar structure, dislocation structures and fracturing are favoured (see chapter V).

Since no xenoliths of the pre-tectonic gneisses have been found in the Naros granitoid it is not possible to describe the effects of M_2 on these rocks. No attempt will be made to define P-T conditions on the basis of the one exposure containing xenoliths of deformed Beenbreek megacrystic granite.

E. POST-NAROS METAMORPHIC EVENT (LATE VELLOORIAN)

The episode described above was followed by the intrusion of the Naros granitoid and subsequent migmatisation. Since there is no evidence of migmatisation prior to this episode it will be assumed that all migmatites in the area can be related to the late Velloorian event which resulted in anatexis of certain marginal parts of the Naros granitoid. The mineral parageneses now present in all formations will be described in this section but it must be stated that it is seldom possible to separate the effects of M_2 and M_3 on the pre-tectonic gneisses. As a working hypothesis it is accepted that rocks throughout the area recrystallised during the Velloorian the only exception being where pre-Beenbreek formations (i.e. charnockitic suite rocks and mafic granolites) are still recognisable.

1. *Jerusalem cordierite-sillimanite-garnet gneiss*

Typical parageneses of rocks of this formation are:

quartz + biotite + sillimanite + cordierite + plagioclase ± microcline

quartz + biotite + sillimanite + cordierite + garnet ± plagioclase ± microcline

They are illustrated in AFM diagrams in Fig. 15.

These gneisses are most suitable for determining metamorphic grade because of their favourable mineralogical composition. They are restricted in volume but fortunately have a fairly wide distribution. One of the most important variations as a whole is the quantity of cordierite seen in thin section. This is especially noticeable when comparing typical Kumian parageneses (containing < 2% cordierite) on the farm Kum Kum with other localities. In nearly

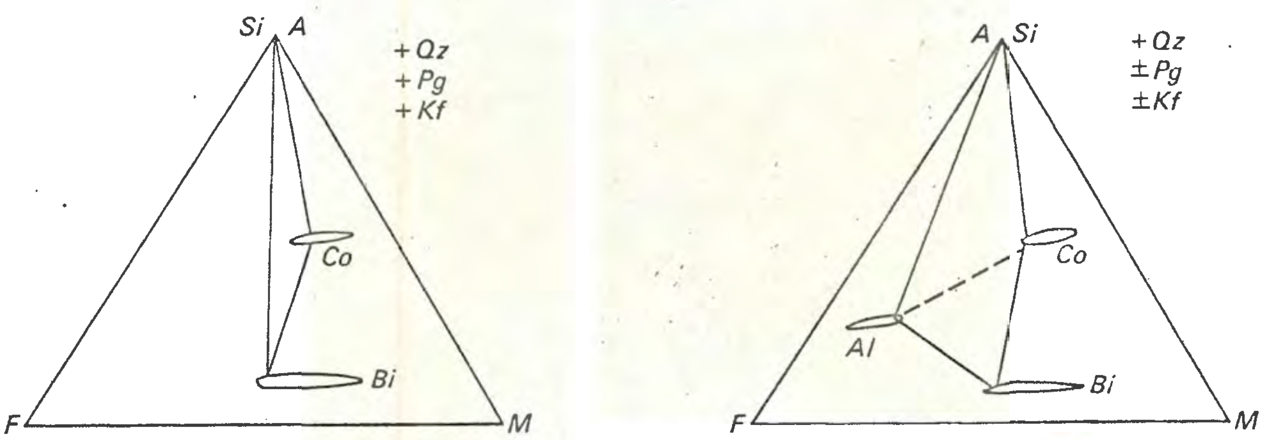


Figure 15. AFM diagrams for rocks of the Jerusalem formation showing Velloorian parageneses. Modified after Reinhardt (1968).

all cases the cordierite shows the same habit described earlier (Section IIIC3), i.e. it forms coronas round garnet crystals, and the amount of cordierite relative to garnet appears to be controlled by the extent to which the reaction *al + si + qz* $\rightleftharpoons$ *cd* has progressed (c.f. Table 3). Some samples taken along the Ham River contain minor amounts of garnet or none at all and it is inferred that the garnet has completely altered to cordierite. Alternatively, cordierite enclosing only biotite and sillimanite may imply the prograde reaction *cl + mu + qz* $\rightleftharpoons$ *cd + bi + si* (Hirschberg and Winkler 1968). Winkler (1974, pp. 89-90) has suggested that in either case a low pressure assemblage is present. In general, the remnant cores of garnet surrounded by an intergrowth of pinitised cordierite, sillimanite and biotite suggests that the cordierite is a retrograde mineral (Section IIIC3). The general absence of primary muscovite which was found in only one thin section, indicates high-grade metamorphic assemblages for the majority of the formation (Winkler 1974, pp.81-86).

On the basis of Jerusalem parageneses the area may be roughly divided into three zones:

(i) high pressure, high or granulite grade, garnet-bearing assemblages in the extreme south west on the farm Kum Kum.

(ii) intermediate pressure, high grade, cordierite + garnet assemblages over the eastern half of the area.

(iii) low pressure, high grade, cordierite (±garnet)-bearing assemblages along the Ham River.

The distribution of these assemblages is shown in Fig. 16 together with data on the distribution of cordierite in the Kambreek formation and the distribution of muscovite in this and the Austerlitz and Orangefall formations.

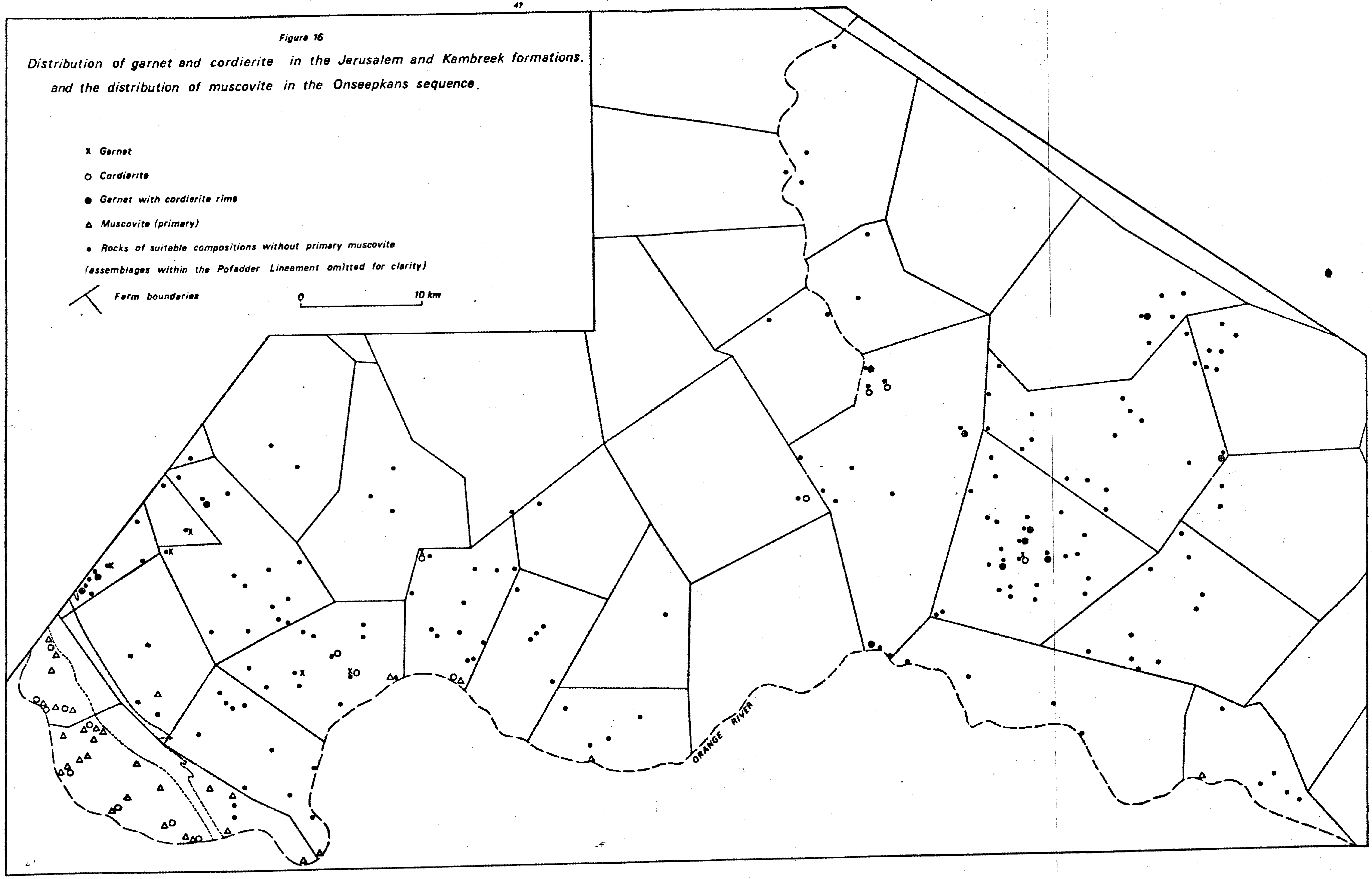


Figure 16

Distribution of garnet and cordierite in the Jerusalem and Kambreek formations, and the distribution of muscovite in the Onseepkans sequence.

2. *Kambreek quartz-muscovite schist.*

South of the Pofadder Lineament occur narrow horizons of lustrous mica schists that are probably the medium-grade equivalent of the high-grade Jerusalem gneisses north of the Lineament (only two horizons of Kambreek schist have been found in the northern block). The most distinctive feature of their mineralogy is the stable co-existence of primary muscovite and quartz. Because no plagioclase is present in these schists the association is not, strictly speaking, typomorphic of medium-grade metamorphism (Winkler 1974, p.81). However, since these rocks are intercalated with quartz-muscovite-plagioclase bearing gneisses their true metamorphic grade is not in doubt.

Typical parageneses are:

muscovite + biotite + quartz ± microcline

muscovite + biotite + sillimanite + quartz ± microcline

muscovite + biotite + sillimanite + cordierite ± microcline

They are illustrated in the AFM diagram in Fig. 17.

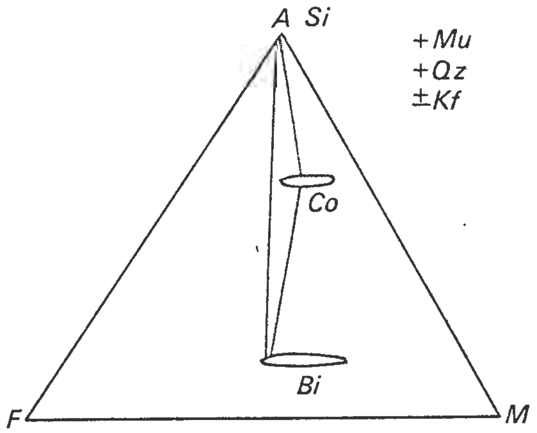


Figure 17. AFM diagram of the rocks of the Kambreek formation showing Velloorian parageneses.

The cordierite in these schists has a completely different habit from that in the Jerusalem gneisses. Here it forms sub-idioblastic grains, commonly unaltered and packed with fine needles of sillimanite and sometimes patches of muscovite and biotite (Fig. 18).

This suggests the prograde reaction *ch + mu + qz* $\rightleftarrows$ *cd + si + bi* (Hirschberg and Winkler 1968). No garnet has been found in these schists, and there is no evidence to suggest that the formation was ever at a higher metamorphic grade than indicated by the present parageneses.

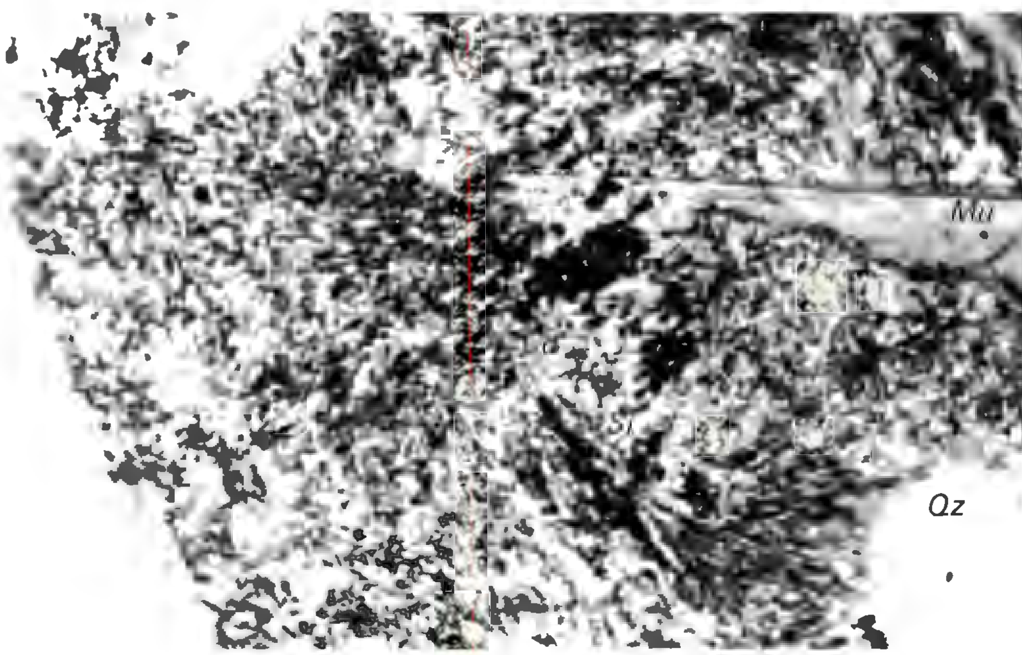


Figure 18. Photomicrograph of a thin section of Kambreek schist showing cordierite and sillimanite forming from muscovite and quartz

3. *Austerlitz quartzo-feldspathic gneiss and Orangefall biotite gneiss.*

Because of the similarity in mineralogy both these formations will be described together. Only one quartzo-feldspathic horizon was found to be intercalated with the mafic granolites and this had a composition free of mafic minerals. Nothing is known, therefore, about the type of Kumian parageneses produced in these rocks. If these formations are interpreted as paragneisses they represent psammites or semipelites and they probably have a high Fe/Mg ratio thus explaining why cordierite is rarely found in these rocks. Because of their composition and restricted mineralogy they are of less use in establishing metamorphic grade.

Typical parageneses are:

quartz + biotite + microcline + plagioclase

quartz + biotite + sillimanite + plagioclase ± microcline

quartz + biotite + garnet ± plagioclase ± microcline

quartz + biotite + sillimanite + garnet ± plagioclase ± microcline

quartz + biotite + muscovite + microcline ± plagioclase

The last paragenesis (medium-grade metamorphites) is only found south of the Pofadder Lineament whereas high-grade metamorphic conditions are defined by the lack of primary muscovite in rocks north of the Lineament. Muscovite is fairly common as a retrograde mineral but not in the sense that it characterises assemblages necessarily typomorphic of lower metamorphic grade. The hydrolysis of microcline to muscovite + quartz is due to a low K^+/H^+ ratio in circulating fluids and the alteration potential of this fluid increases with decreasing temperatures (Hemley 1959).

An interesting feature of the Austerlitz rocks is that garnet is much more prevalent in the east of the area than elsewhere and has apparently formed at the expense of biotite. The mineralogy (but not the texture) is typical of the garnetiferous quartzo-feldspathic gneisses that many authors have described as granulites (c.f. discussion by Behr et al. 1971). Winkler (1974, p.297) and Hensen (1971, Fig. 9) have suggested that almandine forms in high-grade rocks according to the reaction, $bi + si + qz \rightleftarrows al + kf + H_2O$. Although this reaction may explain the disappearance of biotite it has not been observed in thin section and neither does garnet enclose biotite, sillimanite or microcline.

4. Pelgrimsrust hornblende gneiss and schist, Stolzenfels norite, Kum Kum mafic granolite and Jericho amphibolite.

The norites, mafic granolites and amphibolites probably share a common origin and their behaviour during M_3 will be described together. The Pelgrimsrust formation has a similar mineralogy and is also described in this section.

Velloorian retrograde effects on the norites and mafic granolites can best be demonstrated in marginal zones. In the earliest stage a green hornblende can be seen forming coronas around pyroxene crystals and also penetrating the cleavage. Traced towards the contact pyroxene is found in all stages of replacement by a mosaic of hornblende crystals until the norites are finally reduced to amphibolites. In one thin section where primary brown hornblende is present together with the retrograde variety, the former appears to be unaffected by alteration. Velloorian assemblages are prominent in SII and most norites on the farm Beenbreek, but they are only patchily developed elsewhere. Some migmatisation was noted in amphibolites derived from the Kum Kum mafic granolite.

The amphibolitisation of norites and the complete gradation existing between lenticular norite bodies and amphibolite horizons demonstrates the close relationship between the Stolzenfels and Jericho formations. In the east narrow amphibolites preserve all stages of the process and there can be little doubt that the majority, if not all, of the amphibolites is derived from the retrograde metamorphism of the norites. Occasionally rocks of the Jericho formation show no evidence of retrogression and contain no hornblende (DT 630). These rocks are largely confined to the Austerlitz formation on the farm Stolzenfels and may be due to the anhydrous nature of the host rock inhibiting reactions during the Velloorian metamorphic event.

The mineralogy of the Pelgrimsrust formation is fairly simple and consist mainly of hornblende, andesine and quartz. Although this composition is very similar to the amphibolites there is no evidence in the field or from any thin section that the rocks of this formation are derived from the retrogression of other rock types. The hornblende is uniformly green or sometimes blue-green and never brown. Clinopyroxene was found in two samples and showed no hornblende reaction rims, and no relict cores of pyroxene were found as was the case in the amphibolites.

Rocks with this mineralogy offer little help in determining metamorphic grade since they persist from the beginning of medium grade to the beginning of granulite grade with minimal changes in composition. Metabasites at these grades have been described by Binns (1965) in the Broken Hill district, Australia. In the passage from medium to high grade the changes noted by Binns include colour variation in hornblende, disappearance of epidote and the appearance of clinopyroxene. None of these changes are seen in the Pelgrimsrust gneisses but the occasional presence of clinopyroxene may imply high-grade assemblages. This would be in accordance with the grade indicated (c.f. sections IIIE1 and 3) by intercalated horizons of other rock types No orthopyroxene is present thus showing that granulite-grade conditions were not attained. Parageneses for this formation are shown in Fig. 19.

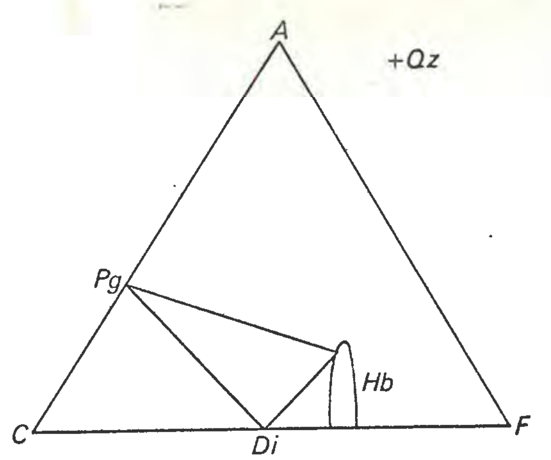


Figure 19. ACF diagram for rocks of the Pelgrimsrust formation.

5. *Keimas calc-silicate gneiss*

Calc-silicate layers occur as narrow horizons intercalated with rocks of the Onseepkans sequence in the west of the area. They have a variable mineralogy but the main constituents are diopside, hornblende, plagioclase and quartz.

Diopside occurs as pale-green equant, sub-idioblastic grains and is found in nearly all samples. Hornblende is usually xenoblastic and has a bright-green colour. It often forms well defined thin seams. Hornblende appears to have reacted with clinopyroxene; the textures are ambiguous but generally suggest that hornblende is the later mineral. A pale-green zoned tremolite was found in one sample. Garnet is not common but when present has a strong body colour indicating a grossular/andradite composition. Colourless garnet from zoned calc-silicate boudins in the biotite gneiss was analysed on the electron microprobe as calcic-almandine. Sphene and epidote are abundant. Scapolite was recorded in one thin section.

These assemblages are not typomorphic of a unique metamorphic grade but the presence of diopside is normally accepted as indicating fairly high grades (Kennedy 1949, Winkler 1974, pp.119-123). Reactions in calc-silicate rocks are not generally temperature-definitive since they are very susceptible to fluctuations in pressure and CO_2 content of the fluid phase. However, certain mineral parageneses seen in thin section afford a rough guide to prevailing conditions especially if the mole fraction of $CO_2$is assumed to vary within normal limits i.e. $XCO_2 > 0,1 < 0,9$ (Metz and Trömsdorf 1968).

In sample DT1358 a pale-green, weakly-zoned tremolite is associated with diopside, plagioclase, quartz, scapolite and trace amounts of calcite. This association is especially significant because, according to Metz (1970), diopside in calc-silicate rocks is derived almost exclusively from the reaction:

$$tr + cc + qz \rightleftarrows di + CO_2 + H_2O$$

This reaction was investigated experimentally by Metz and the stability curve is shown in Fig. 20. The presence of scapolite is apparently not significant. Winkler (1974, p.129) states that it is common at both medium and high grades, and Ramsay and Davidson (1970) believe that its development is due to the presence of halite in the original metasediment.

In sample DT 1955A grandite (grossular-andradite garnet) is found in association with epidote, plagioclase and hornblende. Two generations of epidote appear to be present in this thin section. A yellowish-green epidote (Fe rich?) is found surrounded by a pale-green epidote (Fe poor?) which also completely encloses the grandite. The latter mineral appears to be breaking down to form epidote and ore. In some cases hornblende and yellowish-green epidote are found as cores in grandite crystals which, in turn, are rimmed by pale-green epidote (Fig. 21). A possible reaction producing epidote from garnet is:

$$gr + an + mg \rightleftarrows ep + qz$$

This reaction was investigated by Liou (1973) and the stability curve is shown in Fig. 22. The P-T conditions for the curve agree reasonably well with those given by Metz (1970) for the diopside-producing reaction quoted above and hence the presence of epidote bearing assemblages with diopside bearing assemblages is not inconsistent. The sequence of mineral formation in sample DT 1955A would appear to be hornblende + Fe rich eipidote followed by grandite and Fe poor epidote + ore. Areas in Japan where a prograde metamorphic succession hornblende/epidote is followed by grandite is described by Ito (quoted by Liou, op. cit., p.911). Both prograde and retrograde assemblages are therefore preserved in sample DT 1955A.

The zoned calc-silicate boudins in the biotite gneisses (Orangefall formation) contain a fine-grained association of diopside, garnet, hornblende, sphene, plagioclase and quartz (DT 99A). In the outer zone of the boudins clinopyroxene is broken down to hornblende and the plagioclase is saussuritised.

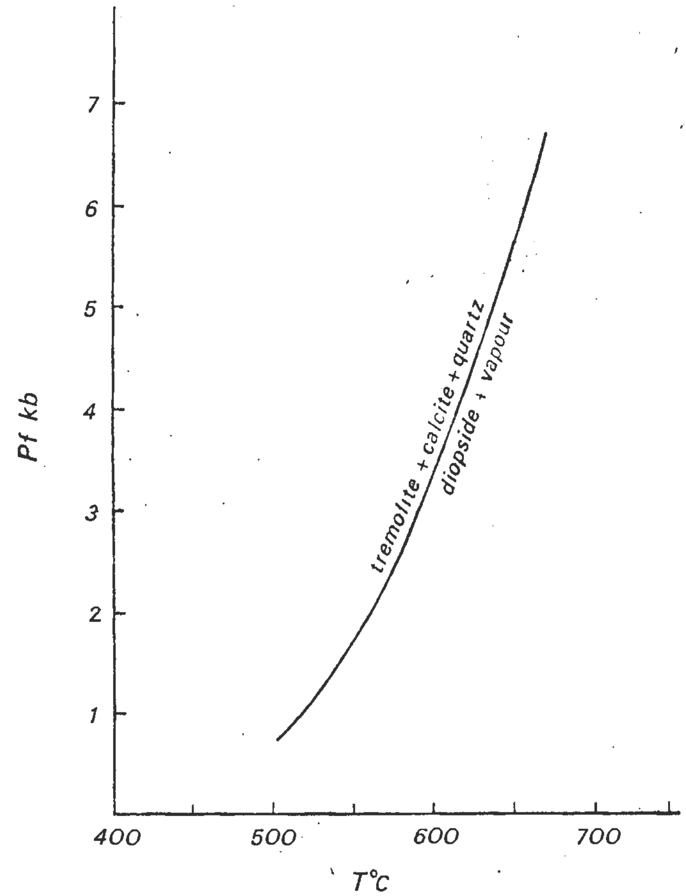


Figure 20. Stability curve for the reaction *tr* + *cc* + *qz* $\rightleftarrows$ *di* + CO_2 + H_2O after Metz 1970, Fig. 15). The curve shown approximates a divariant band.

The garnet is colourless and an electron microprobe analysis showed it to be calcic-almandine (Fe 53%, Ca 31%, Mg 12%, Mn 4%).

The association almandine + clinopyroxene + quartz has attracted attention in recent years following the suggestion by De Waard (1965a) that it results from the prograde metamorphism of orthopyroxene-bearing granolites through the reaction:

op + *pg* $\rightleftarrows$ *cp* + *al* + *qz*

Winkler (1974, p.244) has proposed that, by definition, the association almandine-rich garnet + clinopyroxene + quartz is typomorphic of granulite-grade metamorphism. Up to 20% grossularite component in the almandine is suggested following the constraints of the idealised reaction given by De Waard. However, this figure can only be approximate and will vary according

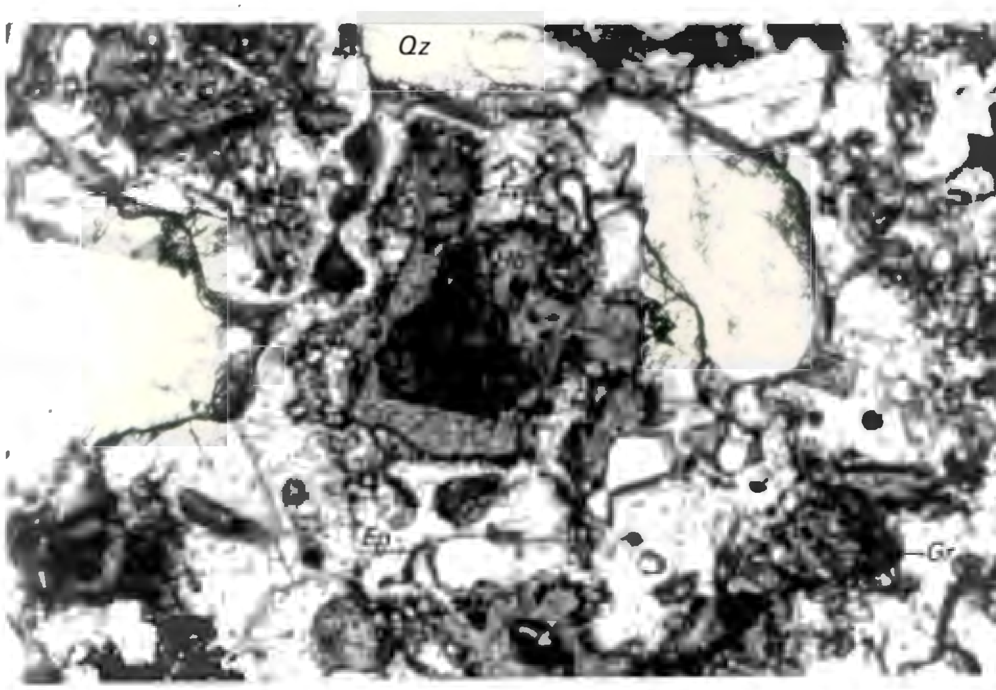


Figure 21. Photomicrograph of thin section DT1955A showing hornblende and Fe rich epidote enclosed by grandite and enclosed in turn by Fe poor epidote.

to the composition of the phases involved. Data on the composition of garnets from granulite-grade terrains are scarce and Winkler's proposal is therefore difficult to assess. Dearnly (1963) quotes analyses from granulite-grade rocks in the Lewisian which all have a high grossularite component and one of his analyses approximates that of sample DT 99A.

In the Onseepkans area there is no supporting evidence for the reaction proposed by De Waard. The breakdown of orthopyroxene to garnet and clinopyroxene has not been seen in any thin section from the mafic granolites nor has orthopyroxene been recorded in any other calc-silicate assemblage.

Calcic-almandine has also been reported from calc-silicate assemblages in the Moinian of the Scottish Highlands which are associated with medium-grade rocks (Winchester 1972). For these assemblages it was suggested that almandine was derived from the reaction:

$$mg + an + qz \rightleftharpoons al \ (Fe\ 67\%,\ Ca\ 33\%).$$

The close similarity of this garnet composition with that of sample DT 99A makes this reaction more acceptable in the present case but no supporting petrographic evidence was found.

The breakdown of clinopyroxene to hornblende in the outer zone of the boudins is undoubtedly due to a later retrogressive event. These boudins occur in high-grade 'quartz + muscovite out' migmatites.

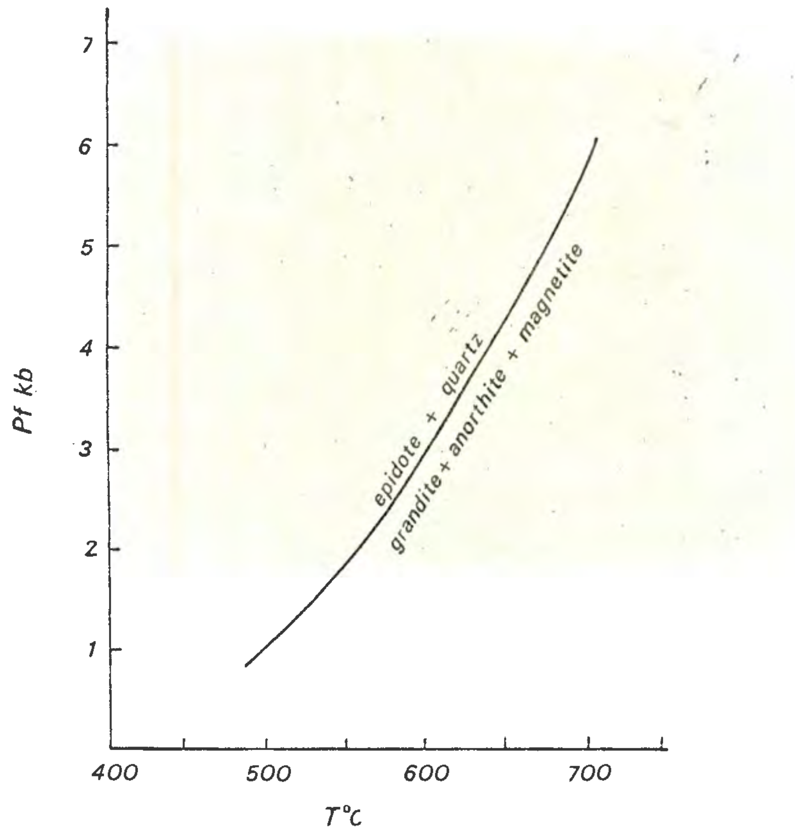


Figure 22. Stability curve for the reaction *gr* + *an* + *mg* $\rightleftharpoons$ *ep* + *qz* after Liou (1973).

In summary, the Keimas formation shows evidence of medium or high-grade metamorphism. No wollastonite has been encountered in calcite-bearing rocks indicating that very high grades were not attained (Greenwood 1967).

Retrograde effects in this formation have resulted in clinopyroxene breaking down to hornblende, grandite breaking down to epidote and, possibly, diopside breaking down to tremolite and calcite. Although it is tempting to relate these changes to the overprinting of Kumian parageneses during the Velloorian event no pre-Beenbreek calc-silicates are available for comparison. Minor prograde and retrograde reactions must be expected in any metamorphic event and without further data the evidence is inconclusive. However, the assemblages do not contradict such an interpretation.

6. *Migmatites*

The migmatites described here are composite rocks featuring leucosomes (leucocratic, newly-formed component) and palaeosome (host rock). They are ubiquitous in the Onseepkans area both north and south of the Pofadder Lineament but are much more common in some formations than in others. The biotite gneisses appear to be particularly prone to migmatisation and where encountered west and northwest of Onseepkans a profusion of leucosomes are found (Plate 15). Blastites are excluded from consideration in this section since they are thought to be a product of pre-Naros metamorphism.

In the east leucosomes often contain garnet but completely lack other mafic minerals (Plate 16). Occasionally these migmatites grade into sticotolites or flecky gneiss (Mehnert 1968, p.37) which are rocks with irregular patches of garnet surrounded by a leucocratic halo depleted in mafic minerals.

There has been a marked variation in the response to migmatisation between the different rock types of the study area. Quartzo-feldspathic gneisses are seldom strongly migmatised, even where they occur as narrow horizons in migmatised biotite gneiss (Plate 17). The margins of the Naros granitoid show various stages in the migmatisation process. In some localities the marginal foliated zone is only a few centimetres wide but leucosomes in the surrounding gneisses clearly penetrate the granitoid (Plate 18a). At a more advanced stage the granitoid margins are penetratively foliated and migmatised (Plate 18b). The ultimate product is a nebulite in which leucosome and palaesome can no longer be distinguished (Mehnert op. cit., p.40) but where the ellipsoidal mafic bodies found in the original rock are still visible (Plate 18c). The development of nebulites is rare in the Naros granitoid but is fairly common in biotite gneiss where it is often accompanied by spectacular fold structures (Plate 19). von Platen (1965) has explained the differing behaviour of rocks during anatexis as a function of their composition. Where the melting behaviour of a single rock type varies temperature or PH_2O fluctuation may be the cause (Winkler 1974, pp.301-306).

P-T Conditions of Velloorian Metamorphism

Two discrete assemblages are present. These are the 'quartz + muscovite out' assemblage north of the Pofadder Lineament and the 'quartz + muscovite in' assemblage south of the Lineament. Anatexis has taken place in both areas but south of the Lineament only cordierite is found in rocks of pelitic composition whereas north of the lineament cordierite is associated with almandine. These data together with the stability curves for calc-silicate assemblages are plotted in Fig. 23.

South of the Lineament the stability field of rocks in this area is limited between the beginning of melting, the 'quartz + muscovite out' curve in the presence of plagioclase (Storre and Karotke 1971) and the lower limit of garnet stability in cordierite bearing assemblages. This is at approximately 650^{o}C between 3,5 and 5 kb.

North of the Lineament the field lies on the high temperature side of the 'quartz + muscovite out' curve (plagioclase present) and in the cordierite plus garnet co-existence field (for rocks with normal Mg/(Mg + Fe))ratios. Calc-silicate assemblages place a restriction on the temperature range of this field. Conditions indicated are 650^{o}-700^{o} C between 5 and 7 kb.

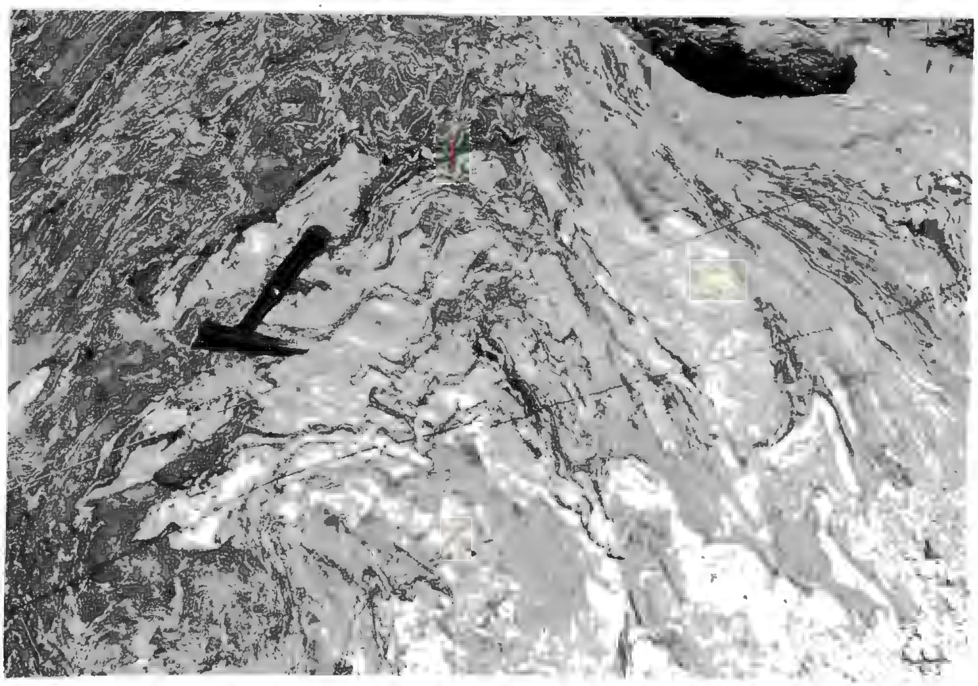


Plate 15. Neosome development in migmatised biotite gneiss (Orangefall formation). Keimasmond Farm.

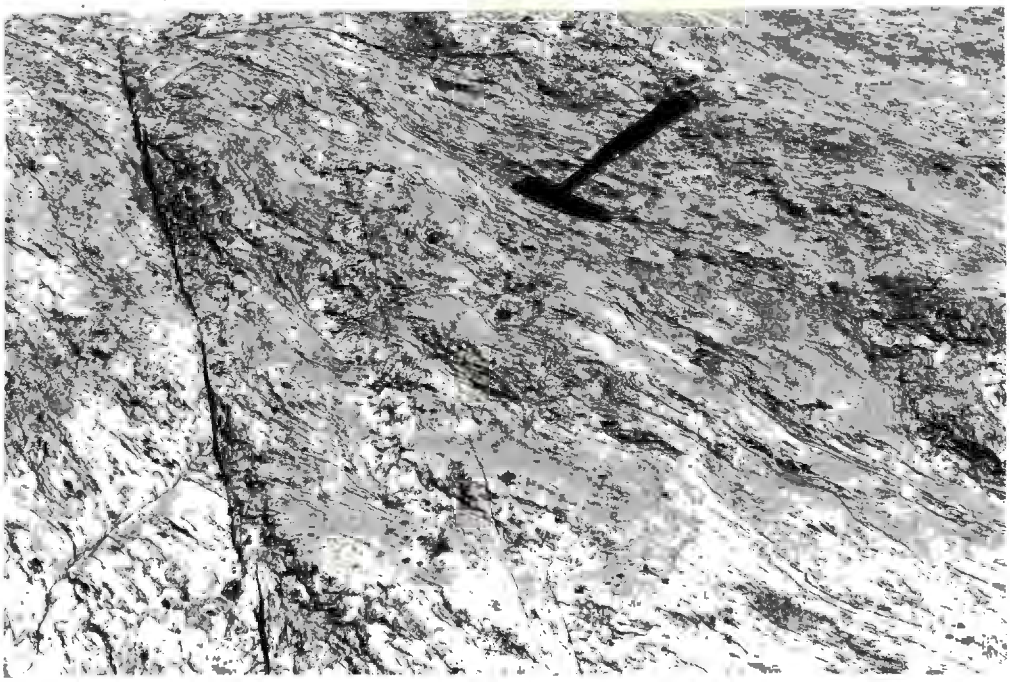


Plate 16. Garnetiferous neosomes in quartzo-feldspathic gneiss (Austerlitz formation). Ondermatje Farm.

Plate 17. Narrow horizons of weakly migmatised quartzo-feldspathic gneiss in migmatised biotite gneiss (Orangefall formation). Keimasmond Farm.

Plate 18a. Initial stage in the migmatisation of the Naros granitoid showing neosomes penetrating undeformed granitoid. Naros Farm.

Plate 18b. Second stage in the migmatisation of the Naros granitoid showing marginal zone foliated and migmatised. Kerelbad Farm.

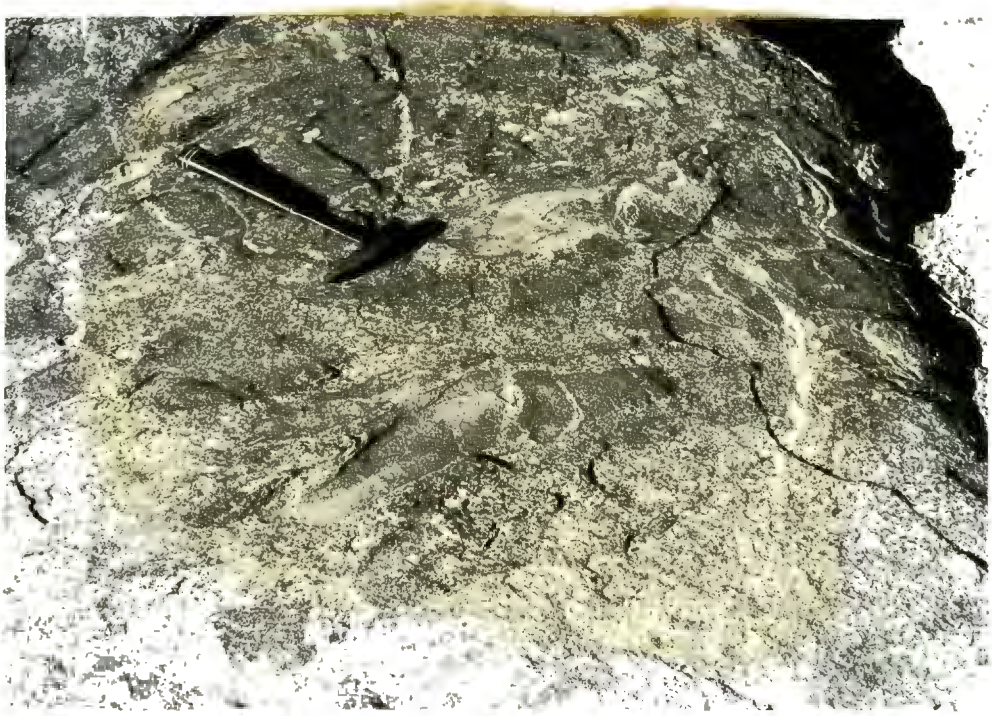


Plate 18c. Completely migmatised Naros granitoid with mafic inclusions still identifiable. Naros Farm.

Plate 19. Nebulite; an advanced stage of migmatisation in biotite gneiss (Orangefall formation). Orangefall Farm.

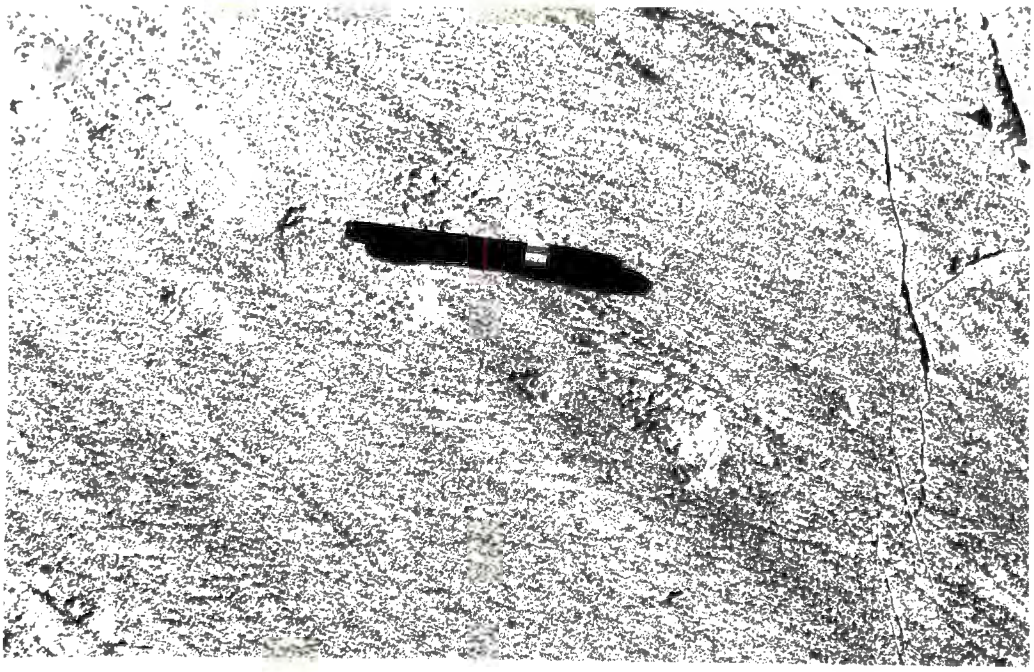


Plate 20. A foliation surface containing two lineations defined by mineral alignment. One trending from upper left to bottom right and the other direction shown by the pen. Jericho Farm.

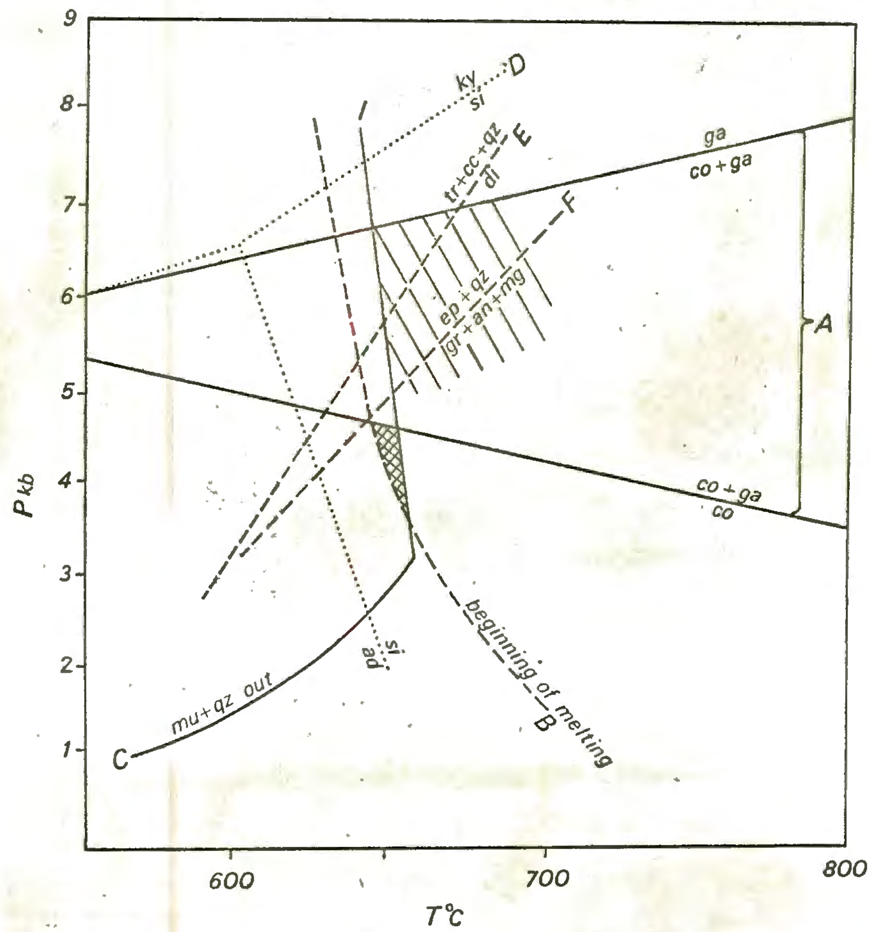


Figure 23. P-T field for Velloorian metamorphism. For reactions see text. Area north of the Pofadder Lineament shown ruled. Area south of the Pofadder Lineament hatched. A. Currie 1971. B. Tuttle and Bowen 1958. C. Althaus 1970. D. Althaus 1967. E. Metz 1970. F. Liou 1973.

F. SUMMARY

A polymetamorphic origin for the Onseepkans gneiss terrain is suggested following the approach outlined in the introduction. The earliest thermal event involved the intrusion of rocks of the charnockitic suite which coincided with a high-grade anhydrous episode of metamorphism. According to Green and Ringwood (1967) and Miyashiro (1973) the mineral parageneses associated with this event are typomorphic of medium-pressure granulite-grade conditions. The products of this early event still exist and are recognisable in a remnant massif centred on the farm Kum Kum. Elsewhere in the area the charnockitic rocks can be positively identified as part of this event but the surrounding gneisses show the effects of later re-metamorphism. The early event is referred to as the Kumian.

Following the Kumian were two episodes of metamorphism accompanied by the intrusion of the Naros and Beenbreek granitoids. These constitute the Velloorian event. The Velloorian culminated in the anatexis of the youngest granite and the stability limit of muscovite + quartz was exceeded north of the Pofadder Lineament. Indication are that migmatisation can be related to this one event thoughout the area. South of the Pofadder Lineament 'quartz + muscovite in' assemblages in suitable rocks indicate a lower metamorphic grade.

Over most of the area the metamorphic mineral assemblages cannot be unequivocally 'pigeonholed' as either Velloorian or Kumian since some rocks have an ambiguous mineralogy and also local fluctuations in grade during both events cannot be ruled out. The choice may be narrowed in many cases by textural evidence. In a large number of thin sections there are indications of a later retrogressive event overprinting mineral parageneses known to be typical of the Kumian. Retrogressive assemblages are not incompatible with the grade of metamorphism attained during the Velloorian. It is suggested that the overprinting therefore post-dates the Beenbreek granite. Overprinting is well developed in the Stolzenfels norites, Kum Kum mafic granolites, Jericho amphibolites, Jerusalem gneisses and, probably in the Keimas calc-silicate gneiss.

One distinction to emerge from the distribution of metamorphic index minerals is the marked change in parageneses across the Pofadder Lineament (Toogood 1975a, 1975b). In the northern block rocks of pelitic to psammitic composition contain very little primary muscovite, and cordierite is frequently associated with garnet, probably as a retrograde mineral. In the southern block primary muscovite has been recorded in all rocks of suitable composition and cordierite is present as a prograde phase without garnet. Kumian parageneses are entirely absent in the southern area and no rocks of the charnockitic suite or mafic granolites or any rocks containing orthopyroxene have been encountered. The most logical explanation for this contrast is that a lower crustal level may be preserved in the northern block (Watson 1973).

A comparison of the metamorphic index map of the Onseepkans area (Fig. 16) with that of the Warmbad area (Beukes 1973, p.168) reveals that the distribution of cordierite and garnet in pelitic rocks in the Warmbad area is exactly as recorded at Onseepkans, i.e. cordierite + garnet are only present north of the Pofadder Lineament whereas cordierite alone is found in the south. A compilation of the data from both areas is presented in Fig. 23. It is also clear from Beukes' map that hypersthene has only been recorded in one locality south of the Lineament whereas it is extremely common in the northern block. Beukes (op. cit.) gives no interpretation for this contrasting mineralogy and, as far as the writer is aware, does not mention it. According to Beukes (in Blignault et al. 1974) 'quartz + muscovite in' assemblages are found throughout the Warmbad area north and south of the Lineament. From field excursions and further mapping near Warmbad undertaken by the writer the presence of muscovite was confirmed for the area south of the Lineament but in the northern block primary muscovite appears to be rare. These observations accord well with the interpretation of the Onseepkans area.

Granolites have been reported from the Warmbad area (Beukes op. cit, pp. 174-175) and orthopyroxene is present in some gneisses of pelitic composition. This is adequately explained on compositional grounds since sillimanite is not found co-existing with hypersthene indicating that the (MgO + FeO)/Al_2O_3 ratio of the rocks is greater than one (Hensen and Green 1971, 1972). The hypersthene gabbros and norites found in the Warmbad area (Beukes 1973, p.248) have an identical appearance and composition to the rocks of the charnockitic suite at Onseepkans. One of these intrusions (SI) occurs at the boundary of both areas and it is reasonably certain, therefore, that the intrusives described by Beukes are part of the Stolzenfels norite formation. A metamorphic history of the Onseepkans area is presented in Table 13.

Comparison with other areas of the Namaqua belt and elsewhere

Rocks of the charnockitic suite have been reported from Grünau (Blignault et al. 1974) and Aus (Jackson 1974). Both areas are part of the Namaqua belt in South West Africa. The charnockitic rocks and granolites in these areas were previously thought to define a 'central zone' of the mobile belt (Blignault 1973, Blignault et al. op. cit.) However, the subsequent recognition of the 'metagabbroids' of Blignault et al. (op. cit.) as charnockitic rocks has led to the conclusion that their distribution is controlled by displacement on the Pofadder Lineament (Toogood 1975a,b). Samples of the charnockitic rocks from the Grünau area examined by the writer have a very similar composition to the Stolzenfels norites and enderbites at Onseepkans and the granolites in this area also suggest a derivation from norites as proposed at Onseepkans. These rocks pre-date the syn-tectonic granites in the Grünau area and probably those in the Aus area (Jackson pers. comm.) and the existence of a high-to-granulite-grade pre-megacrystic granite metamorphic event is therefore suggested for large parts of the Namaqua belt north of the Pofadder Lineament in South West Africa (c.f. Chapter III). Jackson (1976) has estimated the P-T conditions of the granulite grade event at Aus to be 780°-860°C at 4,5-8 kb. The temperature range is the same as suggested for the Kumian but the stable co-existence of cordierite and garnet in pelitic rocks and olivine with plagioclase in mafic rocks indicate lower pressures for the Aus granulite-grade metamorphic event.

In Namaqualand Joubert (1971, p.61) reported norites that appear to have been recrystallised under very high-grade conditions to produce pyroxene granolites. This relationship is identical to that between the norite and mafic granolite at Onseepkans. In parts of Namaqualand, however, a higher grade of metamorphism is suggested by the development of garnet granolites within the pyroxene granolites (Joubert 1971, pp. 45-46). Hypersthene has been broken down by the reaction $op + pg \rightleftarrows ab + al + qz$ (De Waard 1964, 1965a) placing the assemblage in the high-pressure subzone of granulite grade. In the O'okiep area Clifford and Stumpfl (1975) and Clifford et al. (1975) estimated granulite grade conditions at 6-8 kb and temperatures of up to 1000°C on the basis of the alumina content of orthopyroxene and the pyrope content of garnet. The data on granulite grade metamorphism in the Namaqua belt of the northern Cape Province and southern South West Africa are shown in Fig. 24. Little is known about the granulite-grade metamorphism in the Ai-Ais - Grünau area except that both pyroxenes co-exist in metamorphites which suggest intermediate-pressure granolites. In the Pofadder area Joubert (1974b) mentioned the existence of

TABLE 13

Metamorphic history of the rock formations in the Onseepkans area showing stable index minerals and significant parageneses. Unstable parageneses shown in brackets

Formation	Kumian Event		Velloorian Event		
Naros granitoid	post-Kumian		*anatexis*	*high grade*	North of Pofadder Lineament
Grass River augen gneiss	post-Kumian		si (mu + pg + qz absent)	*high grade*	
Beenbreek granite	post-Kumian		si (mu + pg + qz absent)	*high grade*	
Orangefall biotite gneiss	not recognised		si (mu + pg + qz absent)	*high grade*	
Kambreek quartz-muscovite schist	not recognised		si mu + pg + qz	*medium grade*	
Pelgrimsrust hornblende gneiss and schist	not recognised		di hb	*high or medium grade*	
Austerlitz quartzo-feldspathic gneiss	not recognised		si (mu + pg + qz absent)	*high grade*	
Jericho amphibolite	not recognised		hb	*high or medium grade*	
Kum Kum mafic granolite	stable op + cp (ga absent)	*granulite grade ?*			
Stolzenfels norite	ol unstable	*granulite grade*			
Jerusalem cordierite-	Si + ga	*granulite grade ?*	si + cd + ga	*high grade*	

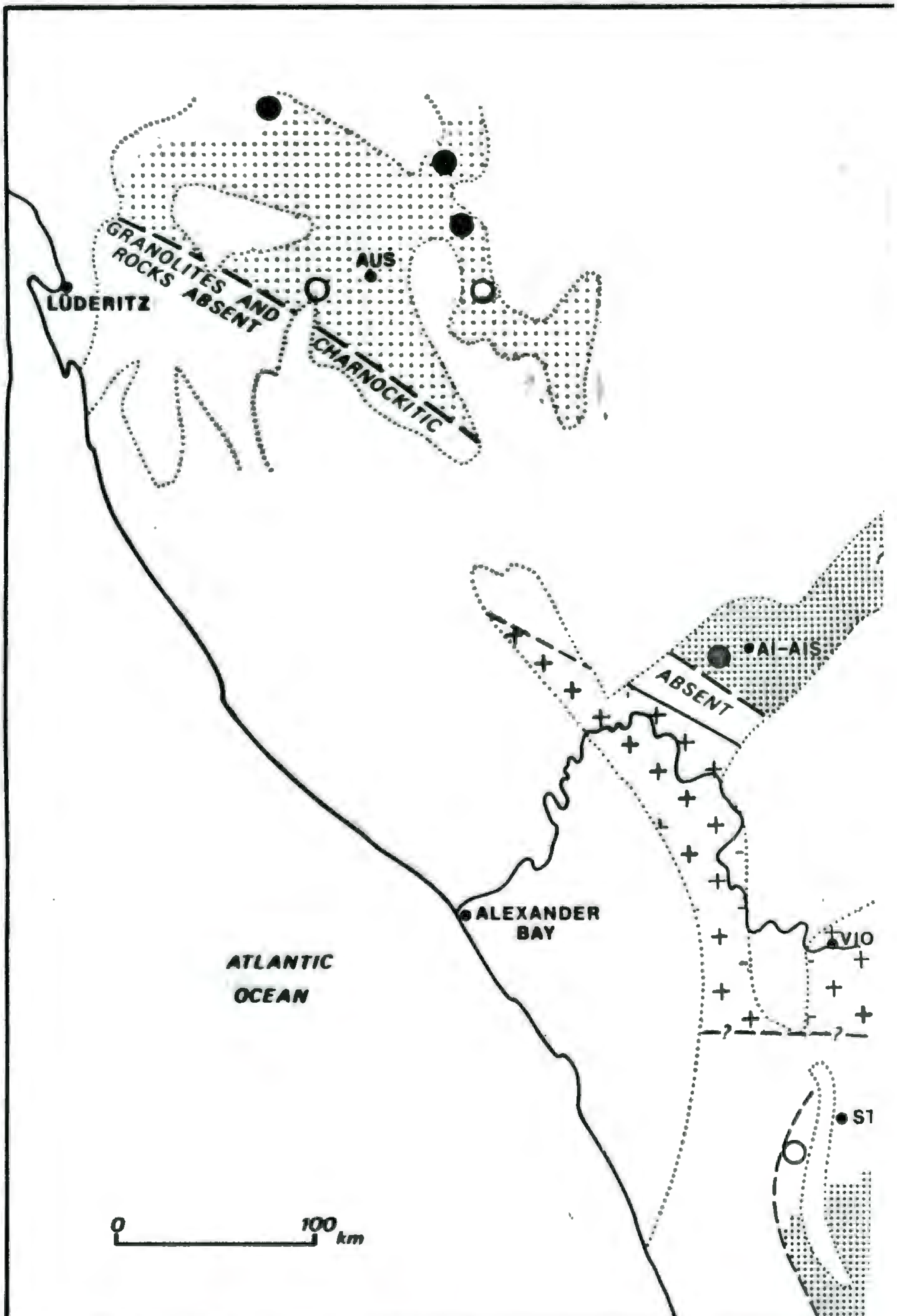


Figure 24. Distribution of charnockitic rocks and granulite-grade metai and southern South West Africa.

granulite-grade metamorphism but gives no details. Intermediate-pressure granolites would be expected however from the distribution pattern seen further to the west (Joubert 1971).

In Namaqualand all the above authors relate granulite-grade metamorphism to the 'main' migmatite-producing event. This is perhaps the most significant difference in interpretation between the Namaqua belt in South West Africa and Namaqualand. It may imply that two granulite-grade metamorphic events have operated in parts of the complex. From work now in progress in the Kakamas area of the Namaqua belt it appears that there is considerable evidence for an early granulite-grade event overprinted by a later medium or high-grade event (Schultz pers. comm.). In a recent publication Joubert (1974b) states that the hypersthene bearing mafic intrusives in Bushmanland predate the granitic gneisses (syn-tectonic granites?). If this is the case it would restrict the area where charnockitic rocks and granolites are part of a later event to the Springbok area of northwestern Namaqualand.

Early high-grade metamorphic event featuring charnockitic rocks in association with granolites have been reported from a number of Precambrian mobile belts including the Lewisian complex in Scotland (Watson 1965), the Grenville Province of Canada (Wynne-Edwards 1969) and the basement complex in Nigeria (Hubbard 1975). A review of these, and other, areas is presented by Crawford and Oliver (1969). As in the present case, these rocks occur as remnant massifs that appear to have survived the effects of reworking during later tectonothermal events.

CHAPTER IV

STRUCTURE – DESCRIPTIVE

The discussion of the structural framework of the area will be dealt with under three headings. This chapter will confine itself to descriptive structural geology; the major folds and fabric elements found in the area and how they are related to one another and an attempt will be made to separate these structures on a time basis. A very well developed shear zone, the Pofadder ZAHNCAFS (Zone of Anomalously-High Non-Coaxially-Accumulating Finite Strain) largely controls the structure of the area west of Onseepkans. Although this could be described in the present chapter it is such a well defined example of rotational deformation that it is better dealt with separately. The last chapter will be concerned solely with the minor folds found in the area, their probable mode of origin, the possibility of using them for making finite strain estimates, and their usefulness in correlation.

A. FABRIC ELEMENTS

A great deal of the following discussion centres around the interpretation of structures and fabric elements found in deformed rocks. Fabric is a term applied to the internal configuration of a rock body and will include such elements as foliation and mineral lineation (Paterson and Weiss 1961). Structure refers to the gross configuration of that body in space and is applied to folds, fault blocks etc. These terms are to some extent dependent on scale and will therefore overlap. Some general comments concerning these structures and fabrics and their notation are given below.

1. *Planar fabrics*

The term planar fabric is used here to describe secondary planar features produced by the alignment of minerals in the rock or sometimes a plane of parting not visibly (in hand specimen) related to mineral orientation. It includes all those fabrics known as schistosity, axial plane cleavage, crenulation cleavage, and gneissose foliation (c.f. Whitten 1966) but excludes flow layering in igneous rocks, joints and faults. Joints and faults are often difficult to distinguish from foliation and are perhaps the greatest source of error in gathering data on planar fabrics. Where the planes of parting (be it apparently 'joints' or 'cleavage') are penetrative, i.e. they appear to pervade the rock at the scale of examination, they are accepted as falling within the definition of foliation. Gneissose foliation is the alternation of felsic and mafic components in a metamorphic rock to produce a banded appearance (Winkler 1974, p. 313). In this sense the term foliation is extended to cover compositional layering due to metamorphism. It shall be of no concern here whether this layering is the result of metamorphic differentiation or derived from partial anatexis (c.f. Hyndman 1972, pp.282-298), but where quartzo-feldspathic *lit* are clearly transgressive they are not included. In practice the problem is not so acute since the constituents in seams of mafic minerals are invariably aligned parallel to the banding. In keeping with standard texts planar fabrics are denoted by the letter s. Subscript numbers, i.e. s_3, refer to the period of deformation during which the fabric was produced.

2. *Linear fabrics*

The term linear fabric is used to describe all secondary linear features. These may vary from the alignment of mineral grains to the elongation of boudins derived from layers several metres in width. The most common linear fabric element is mineral alignment or the segregation of different mineral species to form discontinuous 'trains' on the foliation surface. Occasionally two such lineations may be found on a single surface (Plate 20). A ribbing is also commonly developed on foliation surfaces and this may vary from slight irregularities detectable by running a finger over the surface to large 'macro-ribs' approaching folds (Plate 21).

'Mullion' and 'rods' are to some extent overlapping terms used to describe linear structures produced either by the tendency of the rock to break up into pencil-shaped sections or the linear development of one rock type within another. Wilson (1953) restricted rodding to rocks where the material forming the linear structure was introduced, e.g. quartz veins. This is a useful distinction in lower-grade rocks but becomes difficult to apply in migmatised terrains. Fold mullion occur when the limbs of folded layers are tectonically pinched out during deformation leaving the hinge area isolated in the host rock (Plate 22). Cleavage mullions result from the intersection of a foliation and layering commonly producing linear structures with prismatic cross sections (Plate 23). Mullion structures can be more irregular and it may not be entirely clear whether folding or foliation intersection has exercised a greater influence (Plate 24).

Plate 21. A linear fabric produced by ribbing on a foliation surface. The undulation of this fabric is produced by simple shear. Pelladrift Farm.

Plate 22. Fold mullions represented by isolated fold hinge zones. Nautsis Farm.

Plate 23. Cleavage mullions produced by the intersection of a cleavage with layering. Nautsis Farm.

Plate 24. Irregular mullion structure. Kum Kum Farm.

Plate 25. Lineation produced by the elongation of megacrysts in the Beenbreek granite. Ondermatje Farm.

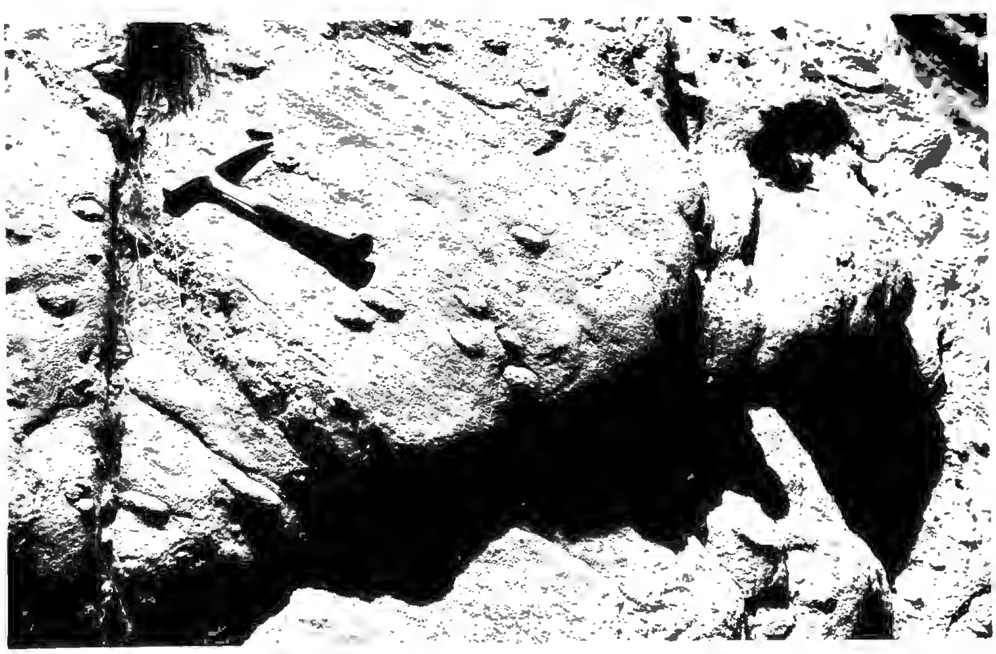


Plate 26. Lineation produced by ellipsoidal sillimanite-muscovite knots in quartzo-feldspathic gneiss (Austerlitz formation). Orangefall Farm.

Boudinage results from the thinning or necking at intervals of a competent horizon during deformation. The segments of rock or boudins sometimes become completely separated cylindrical bodies (Plate 1). One such boudin seen by the writer measured 12 cm in diameter and over 4½ m in length. These boudins should not be confused with a similar-looking end product which results when initially spherical or sub-spherical objects suffer a large constrictive strain. The former are sometimes interpreted as deformed conglomerates (c.f. the pseudo-conglomerates of Ramsay 1956) and as such would provide a very misleading finite strain pattern. A common linear fabric is produced in the Beenbreek granite by the deformation of the megacrysts which gives rise to elongate augen when viewed on foliation surfaces (Plate 25). The deformation of sillimanite-muscovite knots in the Austerlitz formation can also produce a lineation (Plate 26). Very rarely lineations are formed by the intersection of two foliation surfaces.

Lineations are usually annotated l and the accompanying subscript indicates the period of deformation in which it is thought to have formed.

3. *Folds*

Folds are bends in formally planar surfaces. In the Onseepkans area they vary from structures that can be traced for 40 km to folds that are only visible in thin section. It should be noted that the definition of a fold is a geometric one. The definition also requires the pre-fold surface to be planar and it cannot be simply assumed that the surface defining a fold-like structure satisfied that constraint. Fold axes are traditionally annotated B (Whitten 1966).

Several generations of folds have been identified in the study area and their superposition has produced excellent large-scale examples of fold interference patterns. Nine distinct types have been recognised by Ramsay (1967) and since these will be referred to many times in the text they are reproduced here in Fig. 25.

The description and interpretation of minor folds will be dealt with separately in chapter VI and only certain aspects of fold development will be considered in this chapter.

B. APPROACH

There is a wealth of evidence to suggest that the rocks in the Onseepkans area have been deformed more than once. Such evidence includes the interference of folds, or shear zones and folds, the reorientation of tectonic fabrics i.e. foliation and lineation on a mesoscopic and macroscopic scale and the inclusion of deformed rocks within unstrained intrusives.

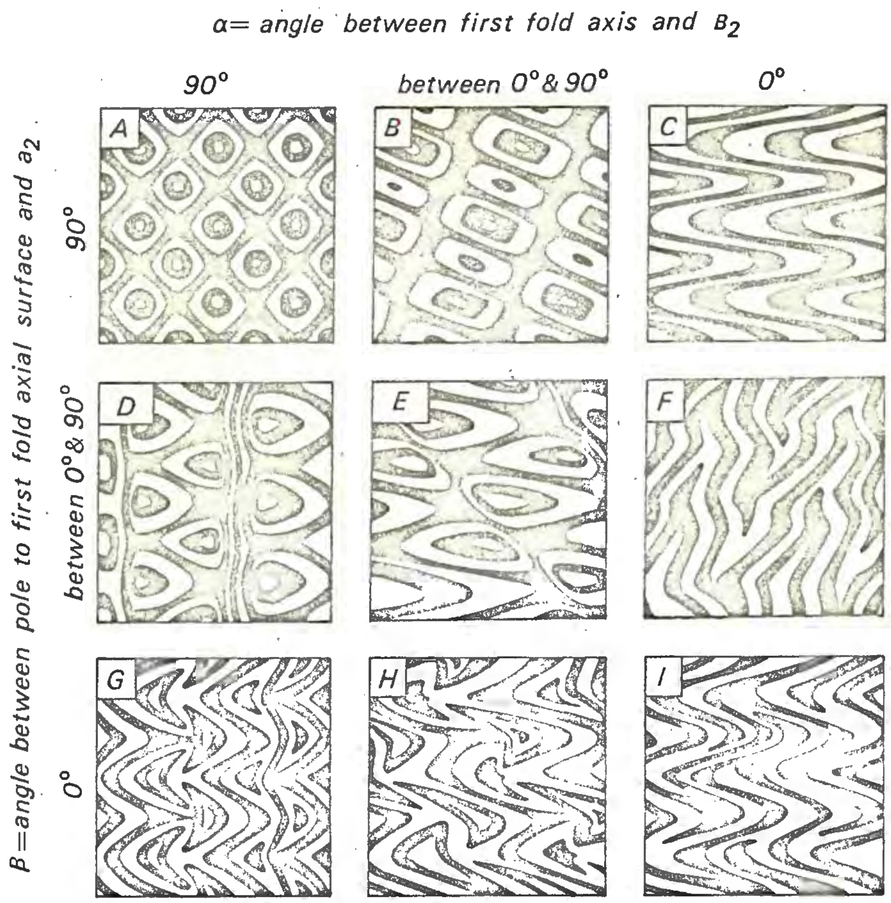


Figure 25. Interference patterns produced by the superposition of two sets of folds. After Ramsay (1967, p.531).

The approach used in this structural analysis has been to identify major folds and shear zones and to classify them into generations or episodes of deformation by observing the effects produced by their interference. Where fabric elements and minor folds can clearly be related to a major structure they too are separated into different generations and are sometimes used to clarify the relationship between major structures.

1. *Data presentation*

Much of the structural data gathered during the investigation is presented on Schmidt equal area orientation diagrams. In all cases the arrow at the margin of the diagram denotes true north. The diagrams have not been contoured since this involves some degree of subjective interpretation. There has been no attempt to define episodes of deformation by the interpretation of the patterns on these diagrams (i.e. 'crossed girdle' patterns). Such techniques of questionable validity since the diagrams in question are seldom subjected to tests on randomness of spatial distribution (Fairbairn 1949, Flinn 1958). In one publication crossed girdle patterns were defined on the basis of a 1% contour representing less than one reading (Naha 1968).

Dots on the orientation diagrams represent poles to foliation surfaces, squares represent poles to fold axial planes, crosses represent lineations and circles represent fold axes. Great circles shown by dashed lines represent fold axial planes (F.A.P.). Great circles representing the distribution of foliation data are shown as solid lines and the pole to this girdle defines the Π axis.

2. *Subdivision of the area*

Rather than treat the whole area in one section there will be a division into sub-areas over which there is a reasonable continuity of individual structures. These sub-areas are not domains *sensu* Turner and Weiss (1963, p.20) but simply areas in which structural data may be discussed without the ambiguities introduced by poor outcrop. Three sub-areas have been defined. These are:

(i) the area east of the contact with the Naros granitoid.

(ii) the area along the Ham River north of the contact with the Naros granitoid.

(iii) the area west of the contact with the Naros granitoid.

The limited exposure at the foot of the escarpment forming the northern contact of the area will be dealt with under categories (i) and (ii) as the case arises. The limits of these sub-areas are shown in Annexure 2. A section at the end of each sub-area discussion will deal with certain kinematic aspects. A fuller description of kinematics is given in chapters V and VI. A summary at the end will outline the possibilities of correlation between these three sub-areas.

C. EASTERN SUB-AREA

Several macroscopic folds up to 15 km in length have been mapped in this sub-area. The interference of some of these folds on the farms Stolzenfels, Jerusalem and Jericho has given rise to elongate dome and basin structures (c.f. Annexure 2). The limb dips on these folds are steep and therefore they are not 'pseudo-folds' produced by the intersection of recumbant-fold crest lines with the earth's surface (Donath and Parker 1964; Weiss 1957, p.51).

The macroscopic structural pattern can be interpreted in terms of two intersecting sets of folds. More than one possible interpretation is presented by the data and Figs. 26 and 27 illustrate the configuration of the fold axial traces for these two alternatives. It is obvious that considerable variation

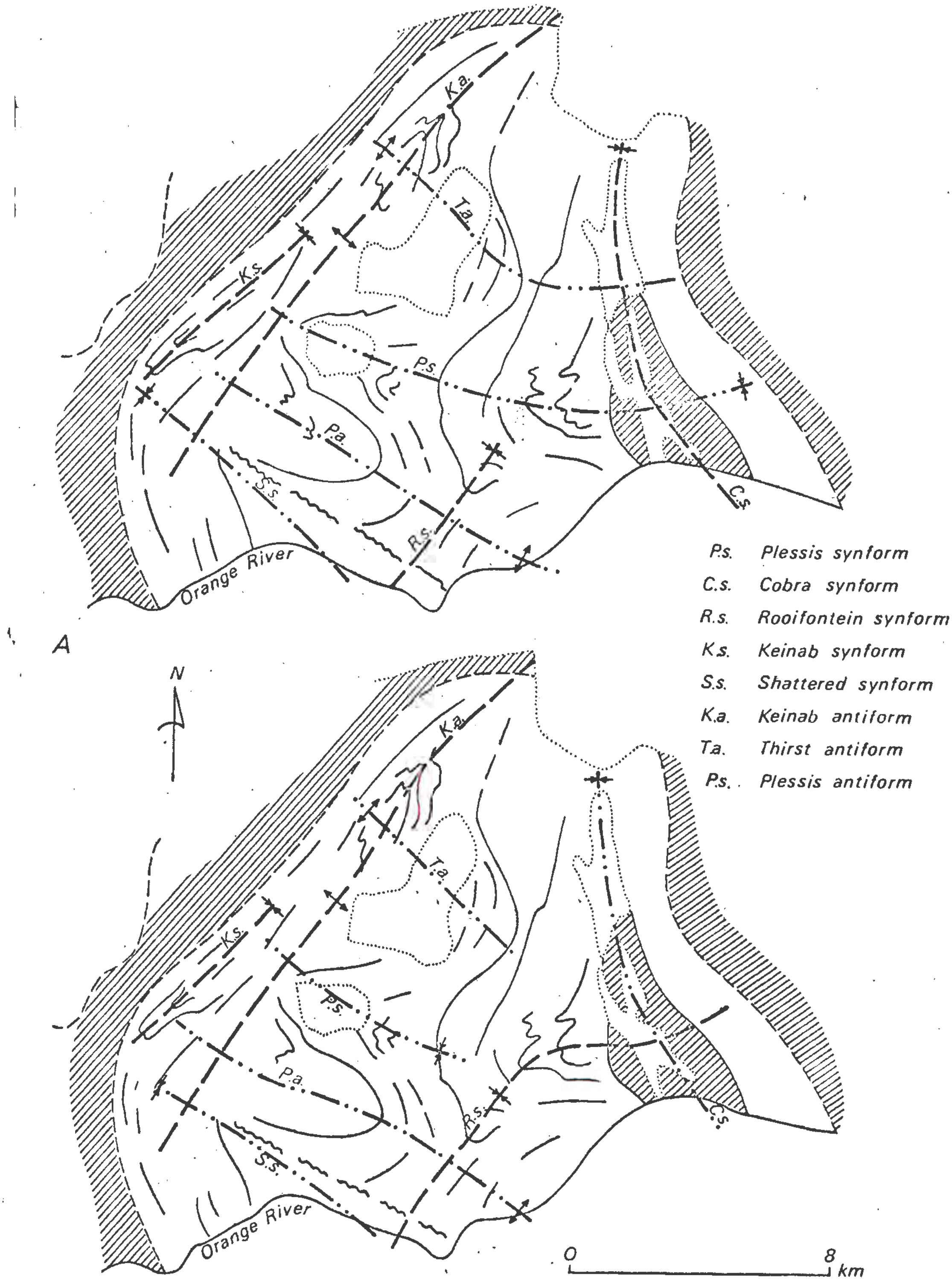


Figure 26. Two possible interpretations for the macroscopic fold pattern seen in part of the eastern sub-area. Ruled area represents outcrop of Beenbreek granite. Blank area are occupied by Austerlitz quartzo-feldspathic gneiss.

in the trend of individual fold axial surfaces exists. The axial trace of the Cobra synform, for example, shows a trend of 135° in the south and a trend of 180° in the north - a difference of 45°. In interpretation B (Fig. 26) the extension of the Rooifontein synform must change in trend from 045° to 085° (a variation of 40°), in interpretation A the extension of the Plessis synform must change in trend from 115° to 085° (a 20° variation). It is this known fluctuation in axial trend that admits various interpretations for this macroscopic fold pattern. It must also be remembered that one of the two major rock types in this area is an intrusive rock (the Beenbreek granite) and does not occupy a single 'stratigraphic' horizon.

The elongation of the domes and basins offers no assistance in this analysis. Their shape is entirely dependent upon the wave length, amplitude and persistence along trend of individual folds. In this respect the idealised fold interference patterns described by Ramsay (1967, pp.520-537), O'Driscoll (1962, 1964) and Turner and Weiss (1963, pp.129-143) in which consistency of elongation is shown are unhelpful (c.f. interference patterns presented by De Sitter 1952).

Since the macroscopic pattern has no unique interpretation, recourse must be made to fabric elements associated with individual folds.

In the hinge zone of the Plessis antiform and the Plessis synform a very strong linear fabric is present plunging at moderate angles to the southeast. This fabric is very well developed in some localities and takes the form of a mineral alignment in the quartzo-feldspathic gneisses (Plate 27) completely obliterating the foliation in the rock. This lineation is unquestionably a B-lineation and it is parallel to the plunge of the major fold axes (Fig. 28). The lineation is strongly related to the degree of amplification of the folds and if followed along the axial trace, it decreases in intensity both westwards towards the Keinab River and eastwards towards the Orange River. The linear fabric may disappear entirely in places and be replaced by a planar fabric which takes the form of planes of parting commonly described as fracture cleavage. This cleavage is parallel to the axial planes of the Plessis structures and cuts an earlier foliation which describes the folds in the quartzo-feldspathic gneiss.

If the linear and planar fabrics associated with the Plessis synform are followed in the direction of plunge a significant change in trend takes place. In the core of the synform the lineation trend is 120° but changes gradually to a more easterly direction as the Cobra synform is approached. Just west of the axial trace of this latter structure the lineation trend is approximately 080°. Immediately across the axial trace of the Cobra synform the trend of the lineation changes abruptly to approximately 010°. The fracture cleavage patterns show a similar abrupt change in strike across the axial trace. Although there remains the possibility that different fabrics are represented here, this is thought to be unlikely as no more than one lineation has been recorded in this locality. The change in lineation trend and fracture cleavage strike is strong evidence for the deformation of the Plessis fabric elements by the Cobra synform and therefore the Plessis and Cobra folds do not belong to the same generation. In the hinge zone of the Rooifontein synform the strong linear fabric in the nose of the Plessis synform is folded which suggests that the Rooifontein and Cobra synforms are associated structures. Interpretation

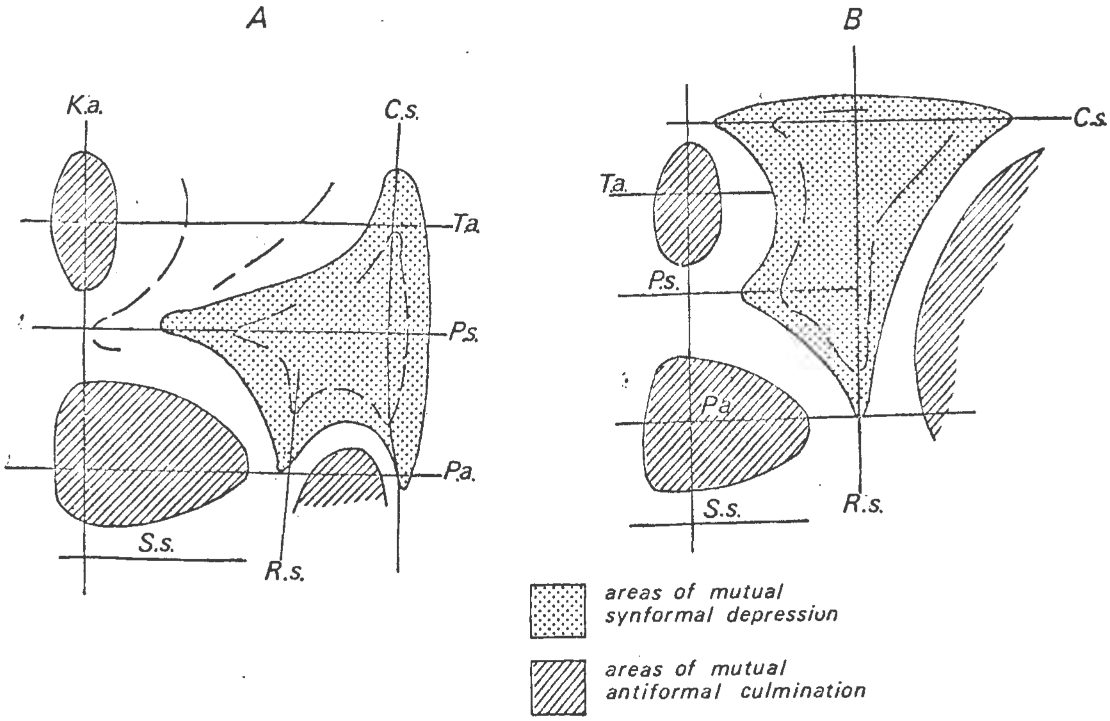


Figure 27. 'Idealised' fold interference patterns assuming the angle between the axial planes of the two fold generations to be 90°. These patterns refer to the interpretations shown in Fig. 26. A and B correspond to A and B in Fig. 26, abbreviations for folds same as in Fig. 31.

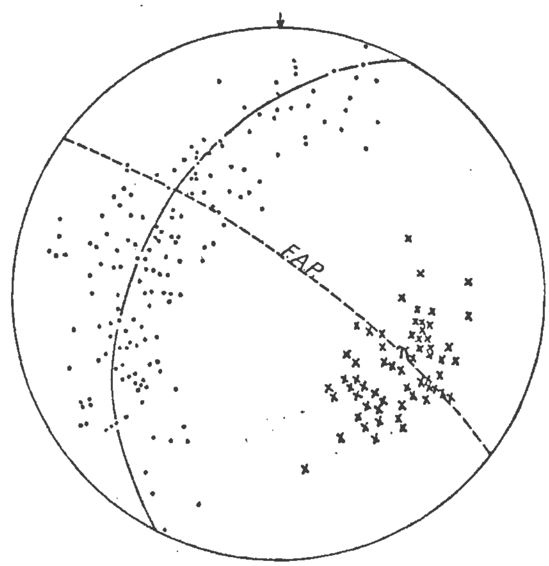


Figure 28. Orientation diagram for lineations and poles to foliation surfaces in a 20 km^2 area in the hinge zone of the Plessis antiform and Plessis synform. 156 poles to foliation surfaces and 59 lineations.

A (Fig. 26) of the above alternatives is therefore favoured. The fact that fabrics associated with one set of folds forming the dome and basin structures is deformed by the other suggests that the folds were not produced contemporaneously in a constrictive strain environment (Ghosh and Ramberg 1968).

The folds comprising the youngest set in this sub-area are the Cobra, Rooifontein and Keinab synforms and the Keinab antiform. The earlier set comprises the Plessis and Thirst antiforms and the Plessis and Shattered synforms. The axial trace of the Shattered synform is marked by a narrow belt of mylonite. When early folds are refolded their axial surfaces may become the site of a major tectonic discontinuity. This results from a rotating shearing action that is caused by, what is in effect, two separate fold axes of the later fold corresponding to the individual limbs of the early structure (Ramsay 1967, pp.546-548).

A notable feature of the younger set of folds is the very wide variation in axial trend that is exhibited over a relatively small area. The difference in trend of the southern section of the Cobra synform and the Keinab folds is about 75° and this change takes place within 15 km. This example strongly supports suggestions by Park (1969), Williams (1970) and Francis (1973, p.185) that fold orientation should not be used as a basis for correlation. Tobisch et al. (1970) on the basis of very detailed work in the Scottish Highlands, have shown that even the trend of young folds may vary by amounts comparable to the present example within 5 km (op. cit., p.256). The assumption by some authors that fold sets are essentially homoaxial over large areas of the crust must consequently be treated with some suspicion (Whitten 1966, pp. 45-48).

1. *Fabric elements associated with the earlier folds*

The strong lineation and the fracture cleavage in the hinge zones of the Plessis folds has already been mentioned (c.f. Fig. 28 and Plate 27). Very few minor folds were found in the hinge zones of the major structures which is probably due to the lack of lithological variation, and hence ductility contrast in the quartzo-feldspathic gneiss. Where mafic horizons are encountered they are often strongly folded but this can seldom be appreciated at outcrop scale. The relationship between the mafic horizons and the quartzo-feldspathic gneisses is not consistent. In the majority of cases these horizons are conformable with the foliation in the gneiss even in the hinge zones of the Plessis folds, which shows that they predate the earliest macroscopic structures encountered in the sub-area. However, in some cases the foliation in the gneiss abuts against the mafic bodies (Fig. 29). This is observed where the mafic rocks have an isotropic fabric and show little or no signs of retrogression to amphibolite facies. This can be an intrusive cross-cutting relationship but it is suggested that it may also be due to the fold forming mechanism (see section IVC4). Where complete retrogression of the norites to amphibolites has taken place a planar fabric is always present and this shows one of three relationships to the surrounding gneisses:

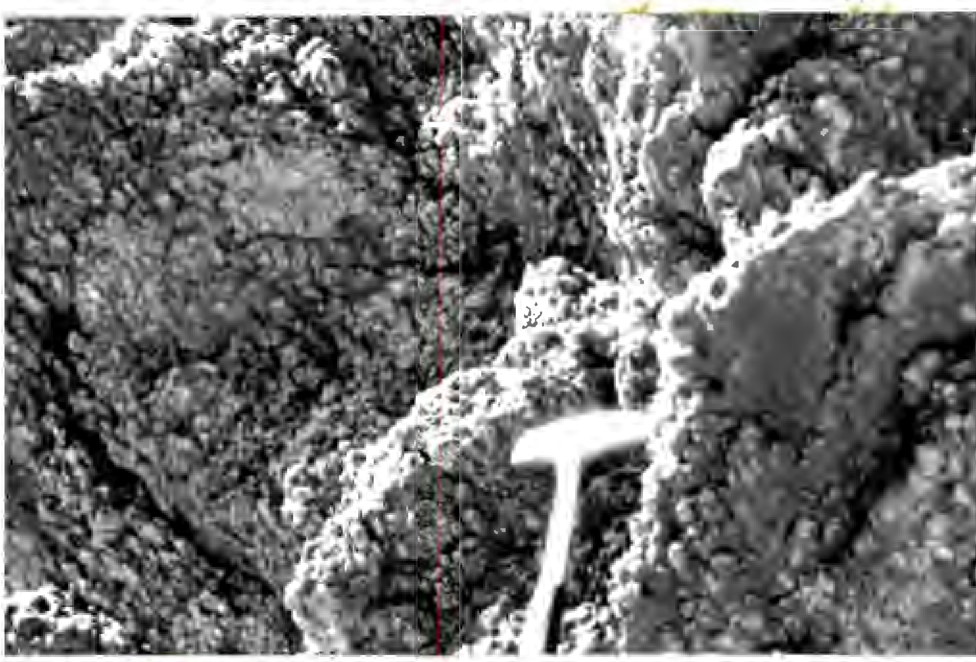


Plate 27. Well developed linear fabric in the hinge zone of the Plessis antiform. Stolzenfels Farm.

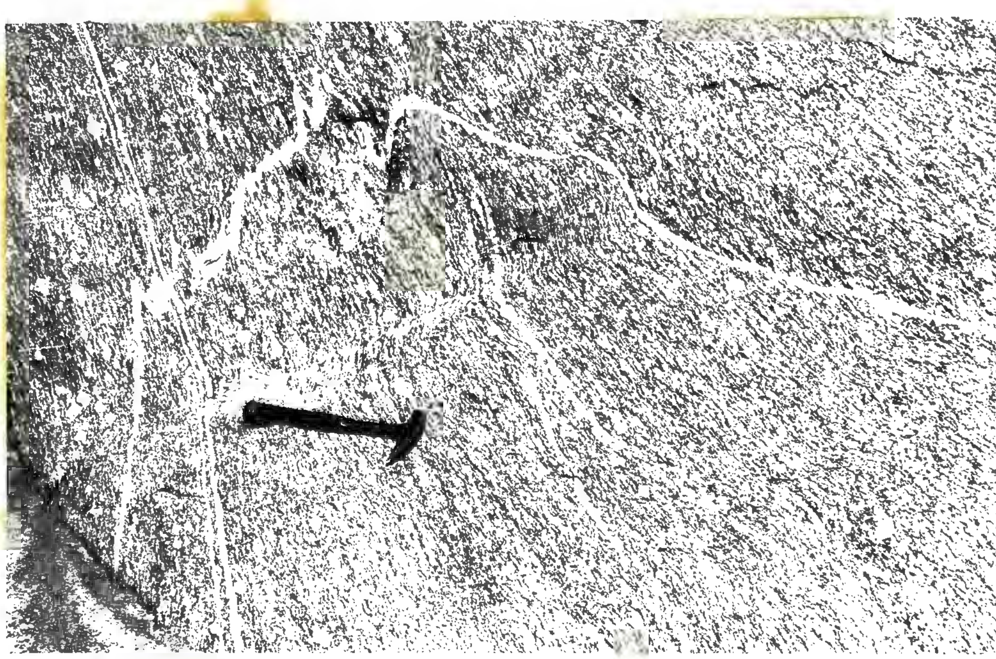


Plate 28. The leucosome on the right hand side appears to be post-tectonic (i.e. intruded into the foliated rock) but on the left it can be seen that the foliation forms an axial plane cleavage on folds defined by the same leucosome. Keimasmond Farm.

(i) the foliation in the amphibolites is parallel to the contacts of the body which are, in turn, parallel to the foliation in the gneisses (Fig. 29c). This is the usual case.

(ii) the foliation in the amphibolites is parallel to the contacts of the horizon which is not parallel to the foliation in the surrounding gneiss (Fig. 29b).

(iii) the foliation in the amphibolite is not parallel to the contacts of the horizon but is parallel to the foliation in the surrounding gneiss (seen in only one case, Fig. 29a)

No penetrative axial plane foliation was found in the amphibolites in the hinges of the Plessis folds and it appears that cases (i) and (iii) reflect a period of penetrative deformation that took place prior to the development of these structures. Case (ii) is more difficult to explain but a possible interpretation is given in section IVC4 (kinematics).

2. *Fabric elements and structures associated with the later folds*

The Cobra synform is the most useful structure for identifying younger fabrics. The deformed Beenbreek granite in the fold core carries abundant leucosomes and remnants of pre-tectonic rocks which have well preserved deformational features. The orientation diagram (Fig. 30A) for the southern end of this synform shows a scatter of poles to fold axial planes along the great circle girdle weakly defined by poles to foliation which probably indicates deformation by the later structure. The exact direction of plunge of the Cobra synform cannot readily be measured in the field and this makes on-site identification of fabrics associated with the fold difficult. There appears to be a preferentially developed linear fabric in the fold hinge defined by the alignment of sillimanite which is approximately parallel to the Π axis obtained from the orientation diagram. The northern end of the Cobra synform (Fig. 30B) in the Beenbreek granite is poorly exposed and the few foliation readings here merely reflect the isoclinal nature of the fold. Some isoclinal mesoscopic folds and a weak foliation are present in this locality and show a plunge to the southeast. These are undoubtedly related to the northern closure of the synform.

On the basis of this fold it can be shown that a hinge-zone lineation is characteristic of both the older and younger structures although the linear fabric associated with the Cobra synform is not as strongly developed as that associated with the Plessis folds. In the core of the Rooifontein synform there is again evidence for a lineation associated with the fold but the predominant linear fabric here is quite obviously related to the adjacent Plessis antiform. This lineation is folded in the synform which has ensured a wide scatter of lineations in this locality (Fig. 31A). All the folds recorded here appear to be related to the Plessis antiform since their axial surfaces show no relationship to the younger synform.

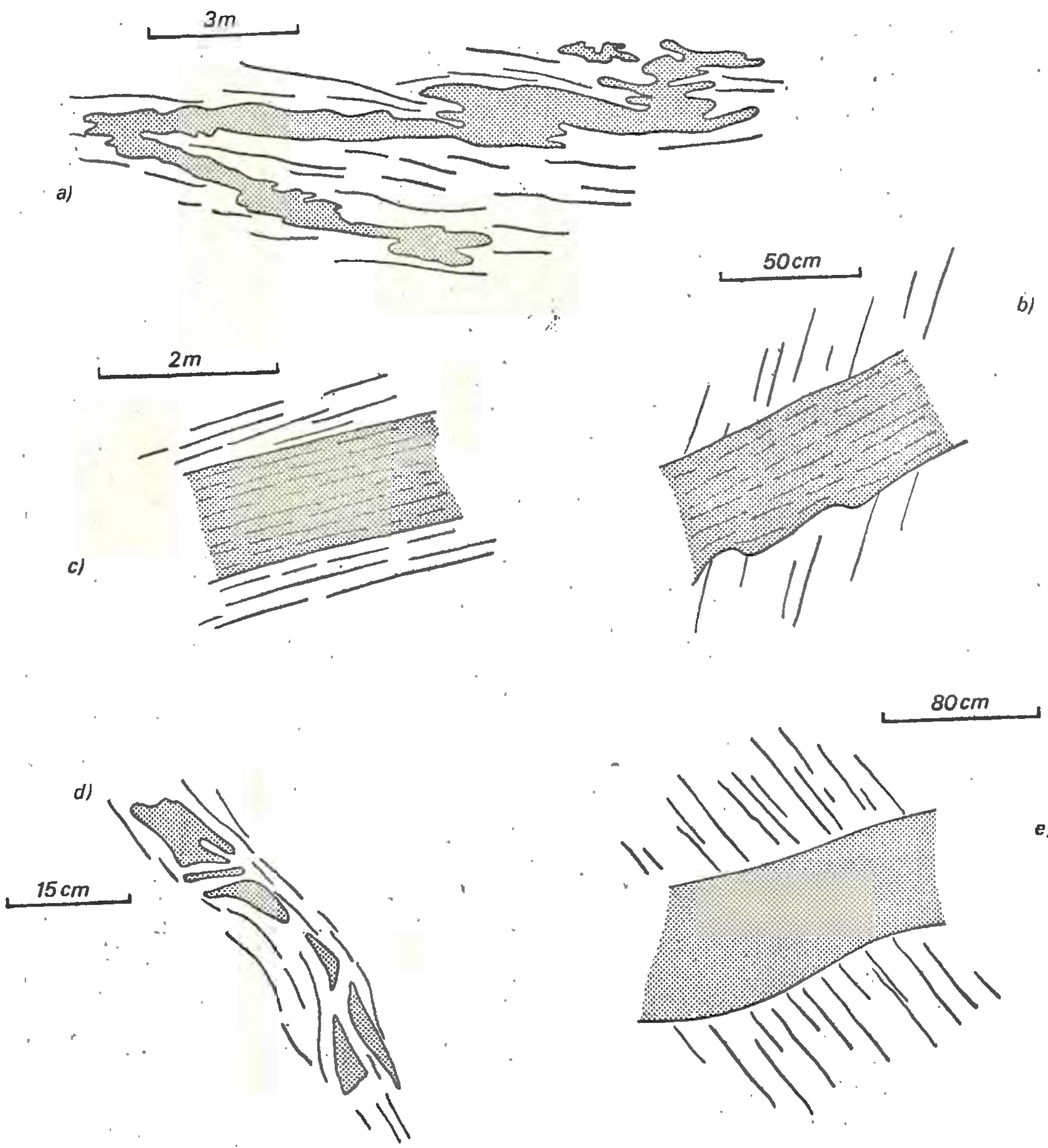


Figure 29. Relationships between mafic horizons and quartzo-feldspathic gneisses in the eastern sub-area.
a) foliation penetrative through mafic rock and gneiss-foliation in mafic rock not parallel to the contacts; b) foliation in mafic rock and gneiss with different orientations - foliation in mafic rock parallel to contact; c) foliation in mafic rock and gneiss parallel and parallel to the contacts; d) unfoliated and boudinaged mafic rock; e) unfoliated mafic rock - foliation in gneiss not parallel to the contact.

Linear fabrics show a small but perceptible change in orientation across the Keinab synform and the scatter of points on the orientation diagram (Fig. 31B) is much smaller. Unlike the two younger synforms described above the trend of the Keinab synform is approximately perpendicular to the older folds and therefore no change in trend of the deformed lineation can be expected across its axial trace. No lineation and no minor folds were found in the hinge zone. The 'bulls eye' plot for foliation data illustrates an isoclinal fold with a sharp hinge.

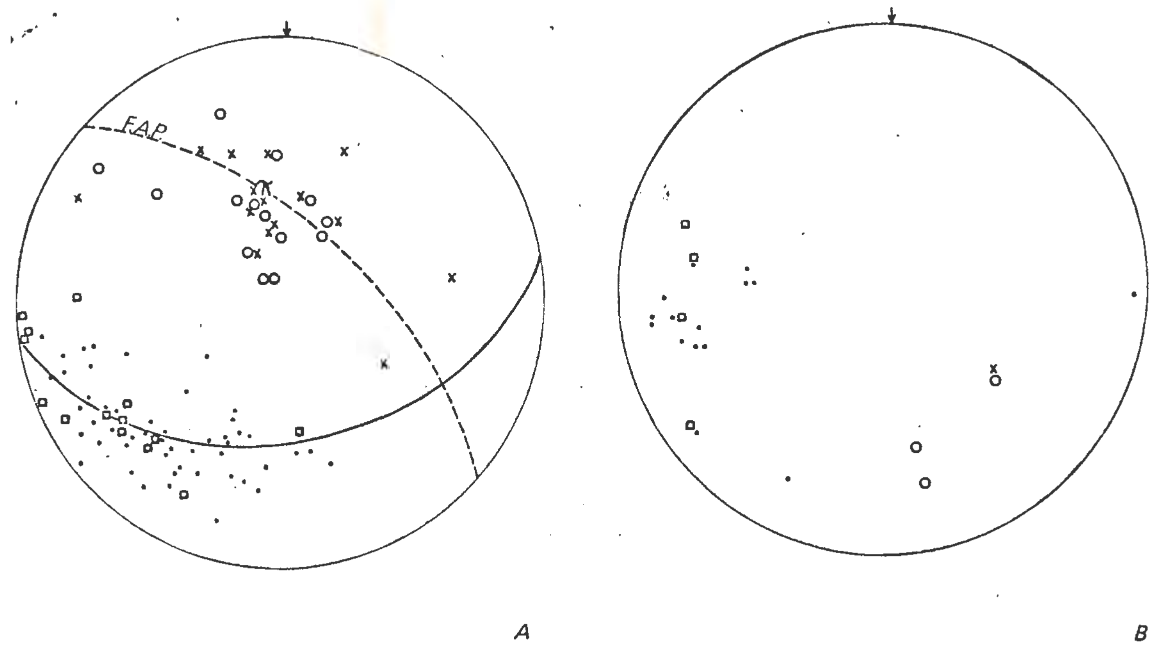


Figure 30. Orientation diagram of data collected from the Cobra synform. A. a 4 km^2 area in Beenbreek granite representing the southern half of the synform. 50 poles to foliation surfaces, 14 poles to fold axial planes, 14 folds axes, 15 lineations. B. a 2 km^2 area in Beenbreek granite representing the northern half of the Cobra synform. 15 poles to foliation surfaces, 4 poles to fold axial planes, 3 fold axes.

3. *Extension of interpretation to the remainder of the sub-area*

If the contact between the Beenbreek granite and the quartzo-feldspathic gneiss on the eastern margin of the Cobra synform is the same as the contact about 1 km to the east (Fig. 26), it is obvious that either a major antiformal fold hinge or a tectonic break must be present in the area. However, no antiformal axial trace could be placed in this strip of quartzo-feldspathic gneiss but the following evidence suggests that it may be present. In the south especially there is a marked change in the lineation orientation from a northerly plunge in the west to a southerly plunge in the east. This change in orientation is similar to that encountered in traversing the Cobra synform and is most easily explained by the presence of a fold. North of the projecting 'tongue' of Nama quartzites, evidence for a fold hinge can be found (point X, Fig. 35), sillimanite-rich horizons and amphibolite in the quartzo-feldspathic gneiss here define the hinge zone of a fold which plunges about 25° to the

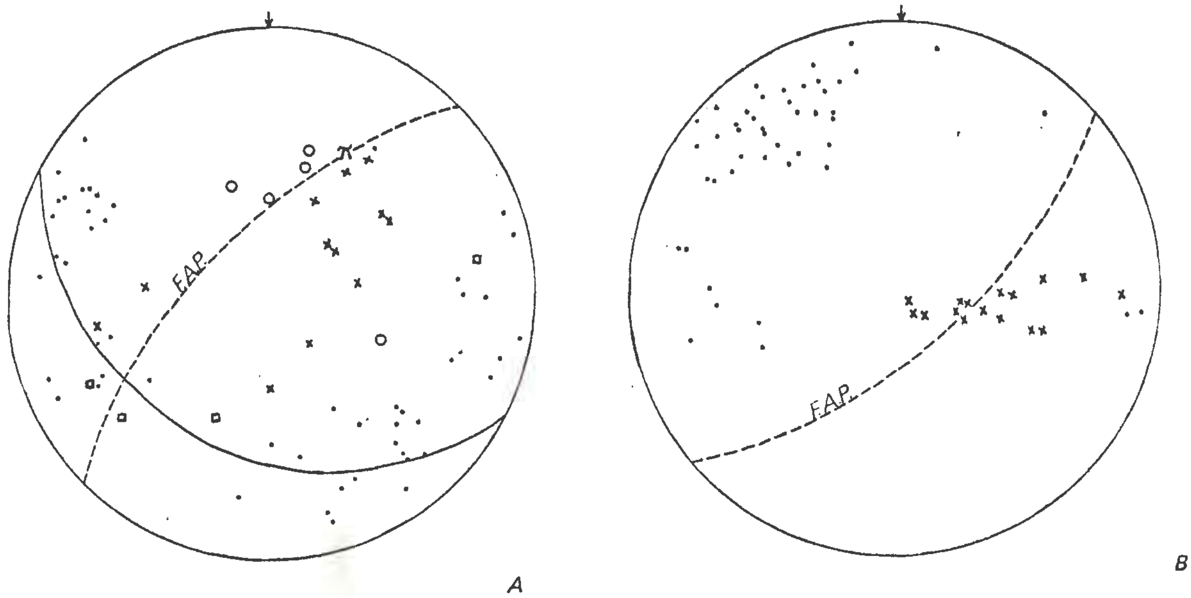


Figure 31. Orientation diagrams of data collected from the Rooifontein and Keinab synforms. A. 9 km^2 area representing the hinge zone of the Rooifontein synform. 50 poles to foliation surfaces, 4 poles to fold axial planes, 5 fold axes, 12 lineations. B. 2 km^2 area representing the hinge zone of the Keinab synform. 46 poles to foliation surfaces, 16 lineations.

northeast (Fig. 32). The lineation, here the alignment of sillimanite and hornblende is strongly related to the plunge of the major structure and does not appear to be the product of an earlier period of deformation. A weakly developed planar fabric is also found in parts of this fold hinge. Minor folds are uncommon. On the basis of this hinge zone and the deformed lineations seen farther south, an antiformal axial trace (the Jericho antiform), is proposed, trending parallel to the Cobra synform axial trace south of this locality to the Orange River. In some outcrops two lineations are present and this corroborates the suggestion of an overprinting of the older linear fabric. Such outcrops may be found along the steep, metalled incline on the road connecting the farms Jericho and Komsberg.

East of the norite body on Stolzenfels Farm (SIV) a large synform is present trending 135° and plunging steeply northwest. The change in orientation of the lineation around this fold is readily seen on the accompanying geological map and on the orientation diagram (Fig. 33). This again suggests that the linear fabric has been deformed by a fold. The axial planes of minor folds have a dispersion similar to the foliation defining the synform (Dead-End synform) which also suggests that they have been refolded. No linear or planar fabrics are developed in the hinge of this synform.

In the extreme east of the area a small elongated dome has been mapped adjacent to the Dead-End synform (see Fig. 35). From the disposition of the lithology it seems reasonable to assume that this structure is the result of two intersecting folds, one approximately concordant with the Dead-End synform

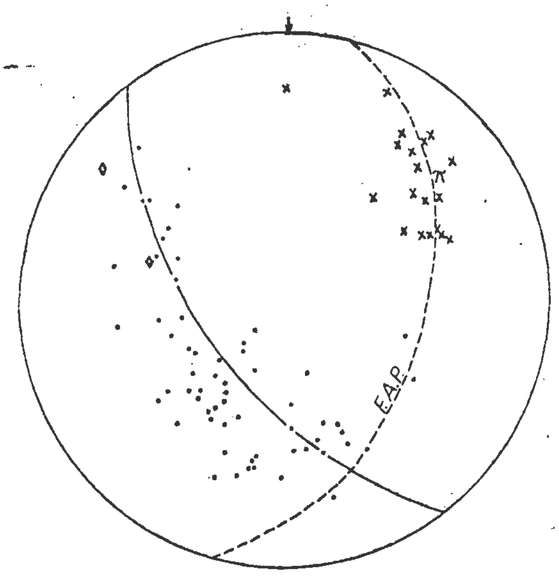


Figure 32. Orientation diagram of data collected from a 1 km^2 area in the hinge zone of the Jericho antiform. 59 poles to foliation surfaces. 2 poles to fold axial planes, 19 lineations.

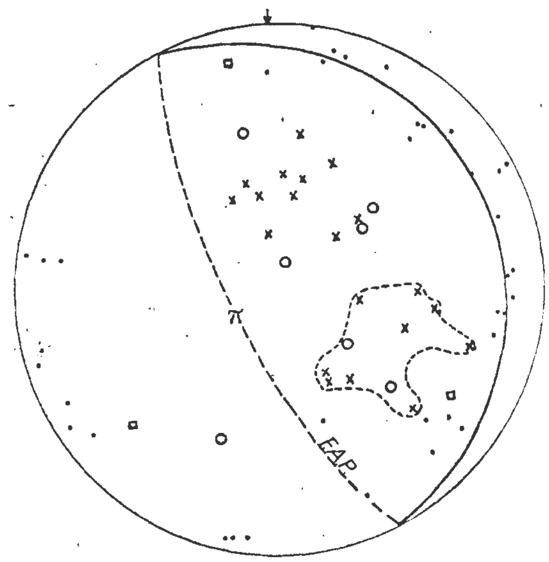


Figure 33. Orientation diagram of data collected from a 55 km^2 area representing the Dead-End synform. 36 poles to foliation surfaces, 3 poles to fold axial planes, 7 fold axes, 20 lineations. Field outlined by dashed line contains fold axes and lineations representing the western limb of the synform

and one trending east-west. Using the interpretation derived above it might be expected that any linear fabric would be parallel to the axis of the older i.e. east-west trending fold. It can be seen from the map, however, that there appears to be some variation in lineation trend across this older fold axial trace. In the south, along the Orange River, the lineation plunges northwest but towards the escarpment, across the axial trace, the plunge is north-northwest. This may indicate that the lineation here is older than either of the two fold generations forming the dome and basin structure in the eastern sub-area. Since both fold directions are defined by a foliation produced by augen-alignment there is clearly evidence for this pre-fold deformation period. However, it may be that this lineation is connected with the older of the two folds but not as a B-lineation, i.e. not parallel to the fold axis (see section IVC4).

Immediately east of the homestead on the farm Jerusalem there is a northeast trending elongate basin. The direction of elongation and general shape of this structure is identical to the dome produced by the intersection of the Keinab and Thirst antiforms 3 km away. The two generations of folds producing this basin are therefore probably the same. The northeast trending axial trace (Goat synform) separates two areas, each with a characteristic lineation pattern. West of the axial trace lineations plunge east but east of the axial trace the plunge is exactly opposed, implying that the linear fabric is deformed about the synform. No linear or planar fabrics have been found associated with the Goat synform. Abundant mesoscopic folds near the centre of the basin appear to be related to the older (pre-Goat synform) fold. These folds show no axial plane foliation but their axes are parallel to the lineation. The Goat synform is probably the major synform east of the Keinab antiform, the Keinab synform being a local, parasitic, structure and there is evidence that its axial trace extends both the northeast and southwest. At the western end of the norite body on the farm Ondermatje the foliation orientation suggests a poorly defined hinge zone which may be part of the Goat synform and to the northeast isolated outcrops near the Keinab River also suggest a continuation of the synform. The axial planes of minor folds in the latter locality have a fairly uniform southeasterly strike which is parallel to the strike of the older folds in the area. However these folds deform a lineation at this locality which therefore suggests the possibility that three lineations of different ages are present in the eastern sub-area. Although the trend of these minor folds is fairly constant it can be seen from the accompanying geological map that their direction of plunge is variable. This may be due to fluctuations in the dip of pre-fold surfaces as illustrated in Fig. 34.

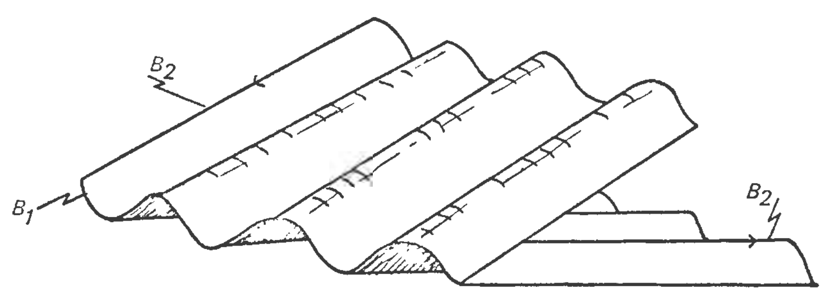


Figure 34. Opposed plunge directions of a second generation of folds (B_2) caused by variation in the orientation of the planar surface related to an earlier fold (B_1).

Using air photos and data from unpublished geological survey maps an interpretation is given below (Fig. 35) of the entire eastern sub-area including the exposed country south of the Orange River. Two sets of macroscopic folds can be identified and it appears that the variability in trend of the younger folds may be due to the presence of large rigid bodies especially the norites and the Naros granitoid. The older fold set has approximately an easterly or south-easterly trend with vertical axial planes. The younger set has a northwest trend and they are predominantly overturned isoclinal folds with axial planes dipping steeply towards the east. It is not possible to give the original direction of plunge of either fold generation since the disposition of the fold axes is largely the result of their mutual interference.

Both sets of folds have associated B-linear fabrics. The older lineation appears to be very common and is strongly developed in the hinge zones of the larger amplitude older folds. The B-lineation of the later folds is found only in some hinge zones. There is evidence from some localities that a linear fabric exists which predates both these fold sets.

The older of the two fold sets deforms a planar tectonic fabric which, in one case, seems to be a penetrative axial plane foliation in folds defined by amphibolite horizons. This foliation may be the same as that developed in the Naros granitoid. Traced outwards from the granitoid the foliation has an identical trend to that in the augen gneisses along the lower reaches of the Ham River. Rocks in this locality are separated from the remainder of the sub-area by a zone of mylonite (the East Arm mylonite belt) which prohibits an unqualified correlation, but the foliation on either side of the zone has an identical orientation. However, a major deformational episode post-dating the intrusion of the Naros granitoid but predating the macroscopic folds which now control the geometry of the eastern sub-area is clearly implied.

It has been demonstrated (c.f. Table 12) that a planar fabric produced by augen alignment pre-dated the intrusion of the Naros granitoid. It has now been shown that an identical planar fabric was produced after the intrusion of the Naros granitoid. This clearly shows that the similarity of such fabrics in the same rock type is not sufficient justification for their correlation.

Neither of the major fold generations has produced a penetrative planar fabric. The older set (typified by the Plessis folds) has a non-penetrative fracture cleavage and the younger set (especially the Jericho antiform) has a weakly developed axial plane cleavage in places. With some minor folds in the augen gneisses an incipient axial plane foliation is produced by the realignment of augen in the fold hinge areas. This only becomes apparent in isoclinal folds and is clearly not penetrative.

Table 14 presents a summary of these conclusions

TABLE 14

Summary of structures and fabric elements in the eastern sub-area

Folds with trend varying from southwest to southeast -weakly developed lineation - weak planar fabrics - mylonites along early fold axial trace
Folds of variable orientation but generally east or southeast - linear fabrics - non-penetrative planar fabrics
Naros granitoid

4. *Aspects of the kinematic interpretation in the eastern sub-area*

The kinematic analysis presented here is confined to a description of the relationship between certain fabric elements and the associated major folds.

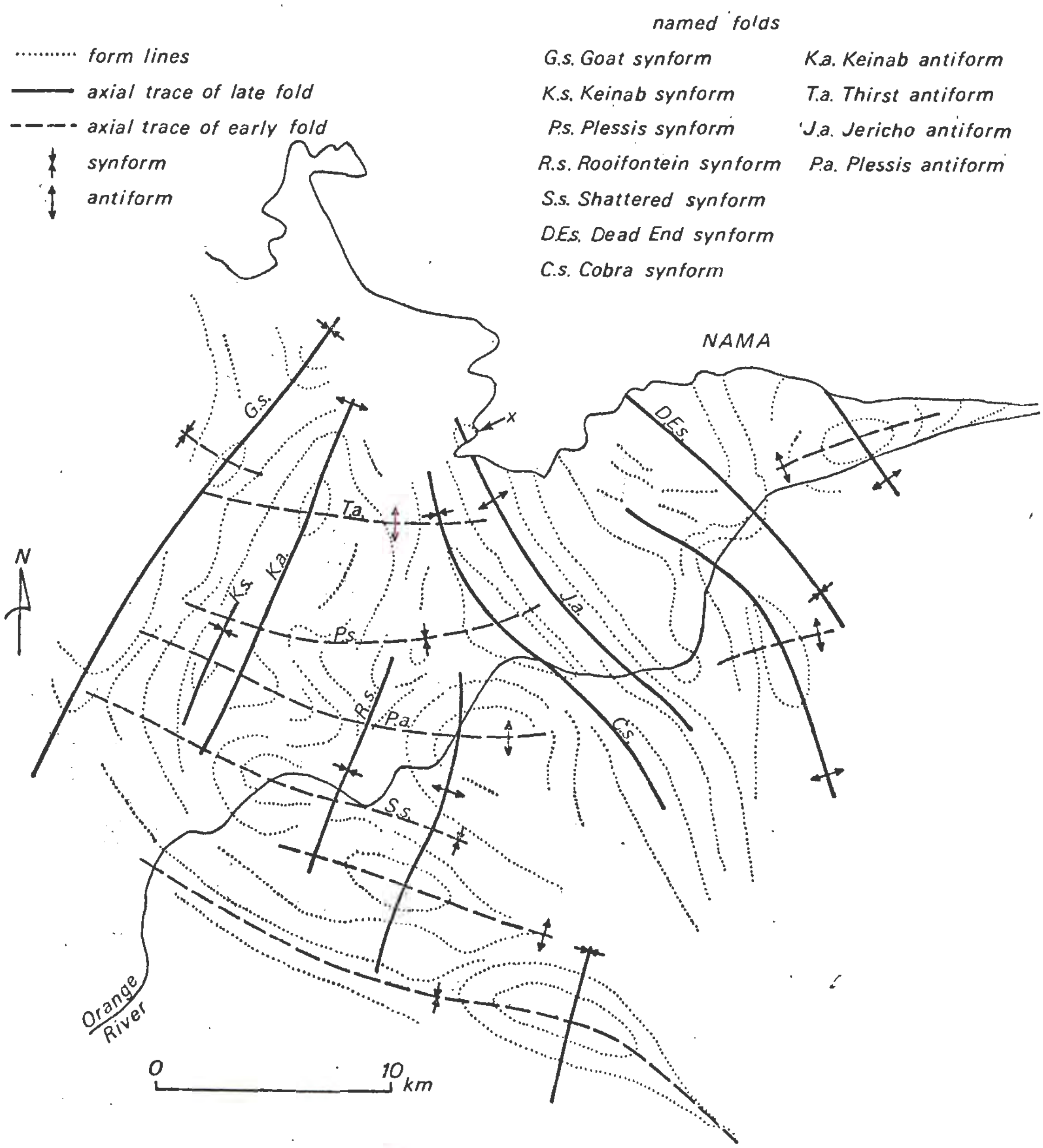


Figure 35. The macroscopic folds of the eastern sub-area. South of the Orange River the interpretation is based on data collected from air photographs and unpublished official maps.

It was stated earlier that there appeared to be a change in the type of fabric elements encountered across the hinge zone of the Plessis antiform.

This variation in style and intensity of the development of the fabric elements can readily be explained by the behaviour of the competent quartzo-feldspathic gneisses during the folding process (Fig. 36). Dieterich and Carter (1969) have shown that there is a significant change in the orientation and magnitude of the principal stresses during the buckle folding of competent rocks. The maximum principal compressive stress (δ_1) increases in amount in the inner arc of the fold relative to the fold limbs and is proportional to the amplitude of the fold. Above the neutral surface the magnitude of δ_1 drops rapidly and it is oriented at right angles to δ_1 below the neutral surface. In isoclinal folds δ_2 will have approximately the same magnitude as δ_1 and δ_3 will be parallel to the fold hinge (op. cit., p. 147). This explains why recrystallization is so localised in many folds and why, as in the case of Plessis structures, there is a marked change in the fabric elements in passing across the hinge area from the inner to the outer arc. The Thirst antiform which is associated with the Plessis structures has a poorly developed fabric which could be explained according to this model by the low amplification of the fold and, therefore, the low magnitude of δ_1.

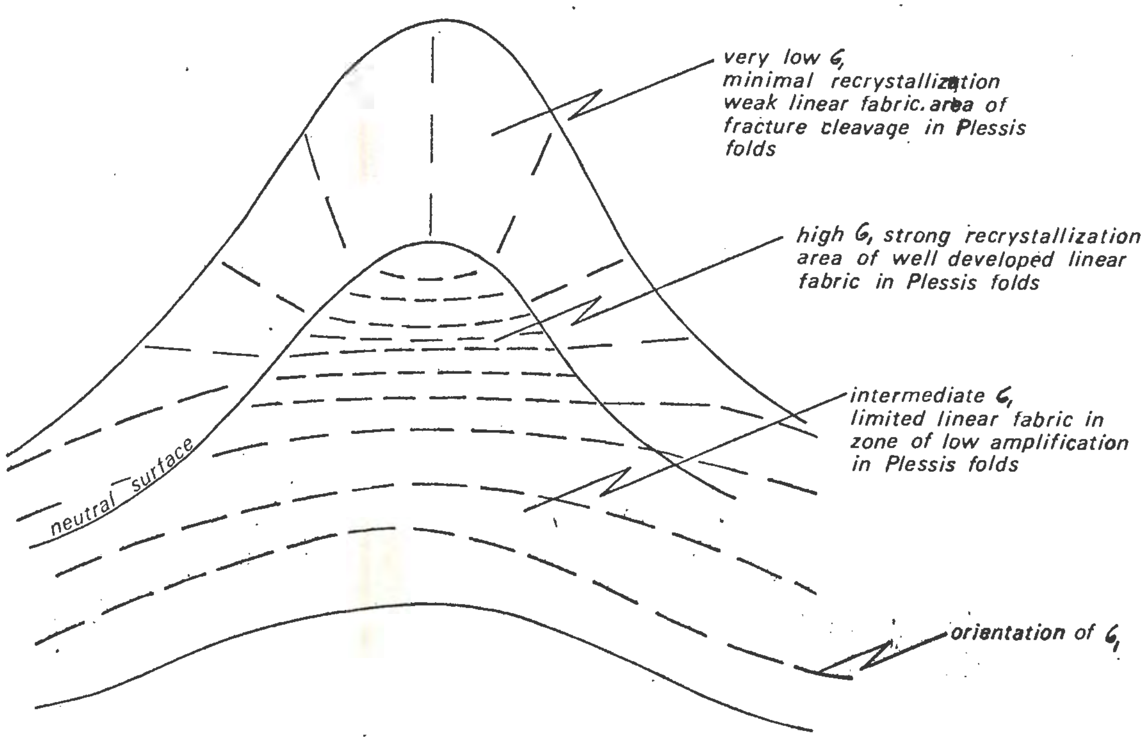


Figure 36. The distribution of stress in the folding of a competent layer by tangential longitudinal strain. Orientation and frequency of dashed lines reflects orientation and magnitude of δ_1. After Dieterich and Carter (1969).

It was also shown in the description of the Plessis folds that the orientation of the planar fabrics in the mafic rocks and the quartzo-feldspathic gneisses was inconsistent (Fig. 29). It is now suggested that this can be understood if the competence contrast between the gneisses and the mafic rocks is taken into account. In most cases it is the mafic horizons that show boudinage indicating that they have behaved more competently than the host rocks (Ramberg 1955). Some retrogressed norite bodies may, however, have deformed in a less competent way than the surrounding gneisses. The effect of changing metamorphic grade on the behaviour of dykes during deformation has been discussed by Francis (1973) and Coward (1973a). They found that the retrogression of a granulite-grade dyke to amphibolite facies resulted in its behaviour as an incompetent horizon relative to the host rocks during deformation. If this is applicable to the Jericho amphibolites certain conclusions can be drawn about their rheological behaviour during the folding process. It is most likely that these horizons have deformed by flexural flow (Ramsay 1967, pp. 391-397) and areas of greater strain will therefore be located on the fold limbs. When the folds approach an isoclinal shape the XY plane of the strain ellipsoid is oriented at a very low angle to the layering and the cleavage can develop almost parallel to the contacts of the horizon (Fig. 37). The foliation in the competent rocks deforming by tangential longitudinal strain (in this case the quartzo-feldspathic gneisses) will still be forming at a high angle to the layering (Ramsay op. cit., pp.403-405 and Fig. 7-83). This process will produce the foliation pattern described in case (ii) above (IIIC*1*).

Unretrogressed norite horizons are probably less ductile than the quartzo-feldspathic gneisses (Coward 1975a, p. 144) and they would therefore fold by a mechanism of tangential longitudinal strain. In this type of fold deformation is restricted to the hinge zones (Ramsay 1967, Fig. 7-63) and the limbs simply rotate. The foliation in the surrounding, more ductile, gneisses will form at a low angle to the norite horizon on the fold limb (Figs. 29e and 37B). This will give the appearance of an intrusive relationship. Similar problems are encountered in migmatites where leucosomes appear to intrude a tectonic fabric (Plate 28) but where the leucosome is folded it can be seen that the fabric forms an axial plane foliation.

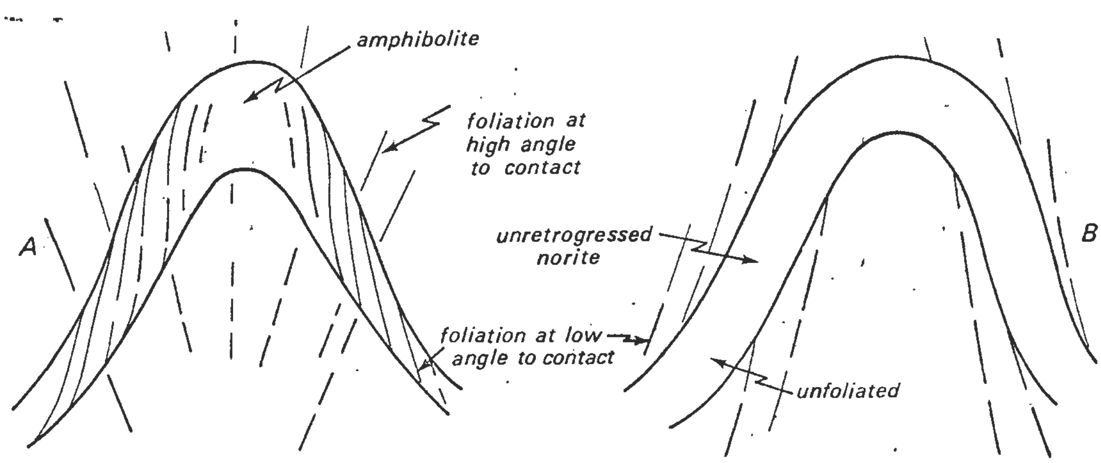


Figure 37. Variation in the relationship of mafic bodies to quartzo-feldspathic gneisses. A. Foliation in mafic rock parallel or sub-parallel to the contacts of the body while foliation in the gneiss is at a high angle to contact. (After Ramsay 1967, Figs. 7-68 and 7-83). B. No foliation in mafic body, foliation in the gneiss at a low angle to contact.

In certain localities on Stolzenfels Farm sub-parallel anastamosing amphibolite dykes in quartzo-feldspathic gneiss show different deformational features. It is not uncommon for some of these dykes to exhibit folding while the others do not and for some to be parallel to the foliation while the others intersect it. In the hinge zones of the Plessis folds amphibolite dykes arallel to the axial planes appear as unfolded linear horizons up to 1 km ɔng intersecting folded amphibolites. These examples might suggest that more ıan one generation of dykes are represented but no field evidence for two enerations was observed. An adequate explanation for this phenomenon can be ɔund in the orientation of these dykes relative to the principal axes of the inite strain ellipsoid. Dykes originally parallel to the axial planes of ıese folds (the XY plane of the ellipsoid) survived the folding process ince they have always remained in the field of extension. Where an unfolded yke is found adjacent to one folded it is proposed that the surface of no-inite-longitudinal-strain lies between them (Talbot 1970), the unfolded dyke eing situated in the extension field and the folded dyke representing the ompression field. This principle is well illustrated in Plate 29 showing vo intersecting neosomes in deformed Beenbreek granite. One neosome is ɔlded while the other, being parallel to the axial plane of the folds, shows no vidence of deformation. Talbot (op. cit., Fig. 1) has found that veins or ykes in the extension field commonly show no evidence of extension such as ecking or boudinage and he has proposed that they have thinned uniformly.

In the dome-shaped structure east of the Dead End synform it was seen that ıe lineation appeared to be unrelated to either of the two fold sets. Although ıis linear fabric may belong to an earlier period of deformation an explanation ın be found in the nature of this fabric element. The augen whose elongation efines the lineation may be envisaged as viscous elliptical particles in a luid matrix. Gay (1968) has shown that under these circumstances the par-icles will not only change shape during deformation (pure shear) but will lso rotate parallel to the long axis of the strain ellipsoid. The orientation f the folds in question with respect to the strain ellipsoid is not known ıt parallelism of the fold axis with the X-direction is perhaps an uncommon ase. It therefore appears quite reasonable to suggest that a single period f deformation will not produce parallelism between the long axes of augen ıd the fold axis.

D. NORTHERN SUB-AREA

This sub-area is restricted to the exposed bed rock approximately follow-g the course of the Ham River as far south as the Naros granitoid.

The dome and basin structures which dominate the geometry of the eastern -area are entirely absent. Seen here are a series of isoclinal folds with ckened hinge zones whose trend is approximately parallel to the Ham River. main folds are from north to south, the Uheib antiform, the Stinking Water iform, the Lost Cat synform, the Bushman antiform and the Ham fold (c.f. ɤure 2). In the southern part of this sub-area the trend of the folds is ɔximately 165° and in the north 110°, a difference of 55°.

Although the dome and basin interference patterns are not seen, the exis-e of other interference patterns may not entirely be ruled out. North he West-Arm mylonite belt as far as the homestead on the farm Ondermatje

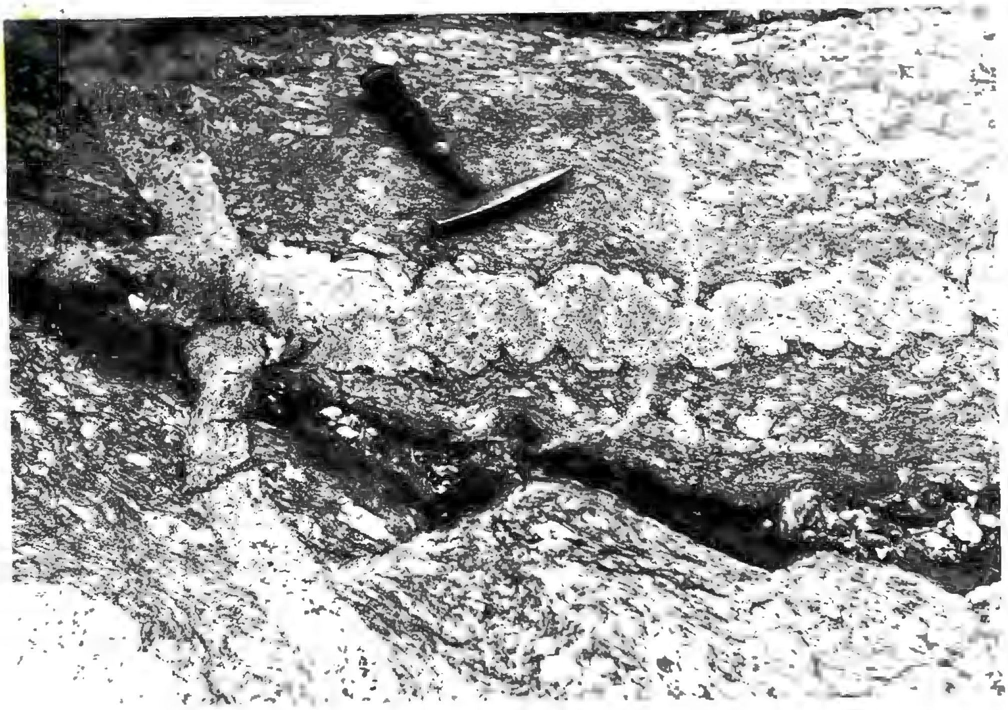


Plate 29. Two contemporaneous neosomes in the Beenbreek megacrystic granite. The one trending from top to bottom of the photograph appears undeformed while the other is folded.

Plate 30. Two generations of isoclinal folds forming a fold interference pattern. The hinge zone of the early fold is seen to the left, and slightly above the clipboard. Velloorsdrift Farm.

(approximately opposite the name Duurdrift Sud 78 - see map) the folds show a fairly consistent trend and plunge north at about 30° (Fig. 38A). There is very little evidence here that more than one generation of macroscopic folds is present. On the eastern limb of the Stinking Water antiform a fold-like structure is defined by foliated Beenbreek granite and augen gneisses. Although the disposition of these formations simulates a fold, a hinge zone could not be conclusively identified and in all probability the pattern results from an irregular intrusion of granite. The interpretation of narrow horizons of metamorphic rocks as isoclinal fold hinges appears to be a fairly common procedure but often the structural evidence is noticeably lacking or not presented (Scotford 1955, Sturt 1961, Goldsmith 1961). This emphasises the necessity of identifying folds on structural, rather than lithological criteria.

Followed northwards it is evident that the homoaxial fold pattern no longer exists. Apart from the obvious change in trend from north-south to northwest-southeast (Fig. 38B) there is also an indication that the direction of plunge changes. Near the homestead on Bokiesbank Ost Farm, the hinge zone of the Ham fold clearly defines a synform plunging steeply to the southeast. This suggests that the Ham fold may belong to a different generation than the northward plunging folds farther south and that type-I (Fig. 25) fold interference pattern may be present. None of the northward plunging folds are, however, actually traceable around the Ham synform and the situation is somewhat complicated by the nature of the Ham synform itself. On the farm Bokiesbank West (approximately opposite the name Ham River on the geological map) the probable continuation of the Ham 'synform' clearly defines an antiform plunging northwest. There is evidence, therefore, that the Ham fold does not only vary in trend but also in plunge and closure which makes it less easy to distinguish from other folds in the sub-area. The fact that both the 'antiformal' and 'synformal' variations of the Ham fold are found in the same northwest trending segment shows that the change in trend is not responsible for this variation. It was suggested in the discussion of the eastern sub-area that the variation in plunge directions of minor folds could be related to the pre-fold geometry, i.e. the orientation of the surfaces resulting from a period of deformation predating the existing folds. This same mechanism may be invoked to explain the plunge fluctuation of the Ham fold. Evidence is given below for the existence of a deformation phase older than that responsible for the macroscopic structures which now exist.

South of the West-Arm mylonite belt the geometry of the rocks appears to be controlled by an antiform (Uheib antiform) trending north-northeast and plunging at moderate or shallow angles to the south-southwest. The continuation of this fold is difficult to place on the map since the local geology is greatly complicated by the intricate association of granitoid and host rock forming the Naros complex. The West-Arm mylonite belt effectively separates this structure from other folds along the Ham River which makes it's relative age uncertain.

1. Fabric elements

Between the West-Arm mylonite belt and the homestead on Ondermatje Farm a fairly clear picture is obtained of the relationship between the macroscopic and the fabric elements.

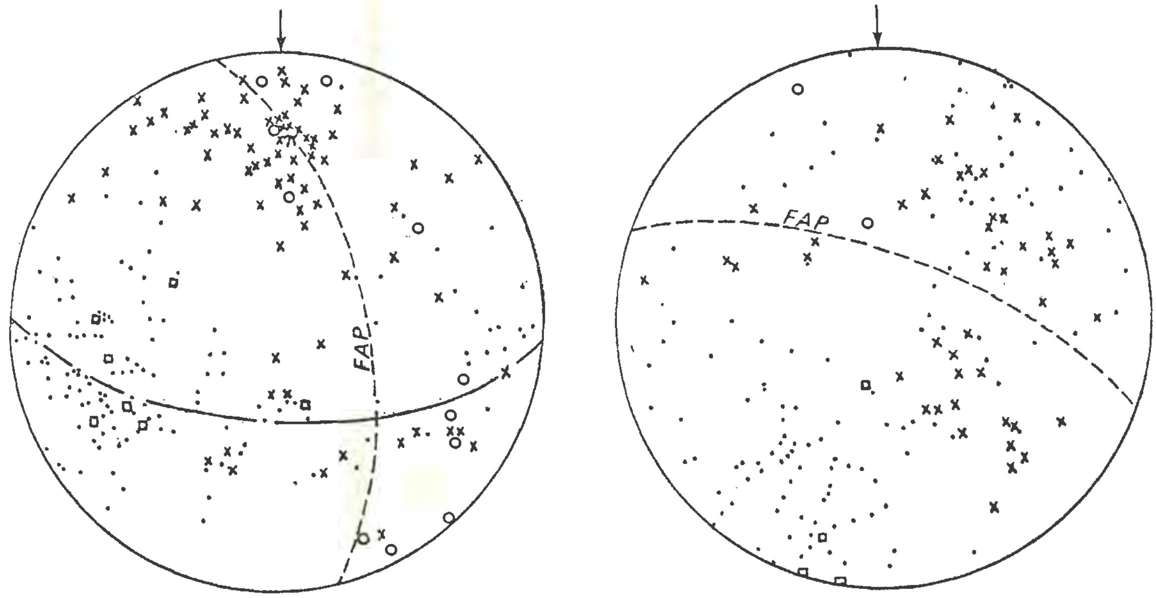
None of the large structures in this area has an associated planar fabric but a lineation defined by mineral alignment and augen elongation is conspicuous in some hinge zones (especially the Stinking Water antiform) and is quite obviously related to the plunge of the folds. In Fig. 38A the Π axis shows a strong relationship to these hinge zone lineations. Many of the lineations on the limbs of these folds do not show such a clear retionship. There are indications here that these folds deform an older linear fabric as was seen in many of the macroscopic folds in the eastern sub-area. The lineation on fold limbs plunges fairly consistently north on the western limbs of antiforms and east or southeast on the eastern limbs. The wide scatter of lineations in Fig. 38A reflects this deformation.

Considering the fact that two linear fabrics must be present, one associated with the macroscopic folds and one which predates them, there are surprisingly few outcrops where two lineations could be recorded. It has been noted by Lisle (1974) that in most areas of basement rocks only one lineation is normally found which has been interpreted by many authors as the 'over printing' of the earlier lineation by a later fabric. However, Lisle (op. cit.) points out that if lineations do not behave as passive markers during later deformation they will be rotated according to the orientation and magnitude of the principle strains during the later event. This mechanism is attractive in that it explains the absence of two linear fabrics in this locality.

To clarify the lineation problem techniques offered by Ramsay (1963) and Elliot (1965, 1968) to rationalise complex lineation patterns were applied. These techniques involve the plotting of lineations as a function of their pitch on the foliation surface or the angle they subtend with later folds. No meaningful results were obtained from these methods and this is probably due to the insufficient data since Elliot (1968, p. 171) mentions data density in the order of 10^2 readings per 1 km^2. This detail could not be attained in the present investigation. Ghosh (1974) has also pointed out that pre-fold layer-parallel shortening results in the deformation of lineations and produces more complex patterns than would be expected from folding alone.

South of the West-Arm mylonite belt the Uheib antiform appears to deform a linear fabric in the manner described before. There is also evidence for a linear fabric parallel to the plunge of the fold (Fig. 39).

The relationship between the Naros granitoid and other formations can better be observed here than on the eastern side of the body. In many places the granitoid intrudes other pre- and syn-tectonic gneisses and there can be little doubt that the foliation defining the Uheib antiform is not the same as that found in the granite. In some outcrops, however, a single cleavage is seen to be penetrative through the granitoid and host rocks demonstrating that both the granite and the surrounding rocks have been sporadically refoliated. There is no evidence in the field or in thin section that the mineral



A B

Figure 38. A. Data collected from an 80 km^2 area between the West-Arm mylonite belt to approximately the position of the name Durrdrift Sud on the geological map. 132 poles to foliation surfaces. 7 poles to axial planes, 11 fold axes, 73 lineations. B. Data collected from a 240 km^2 area along the Ham River north of the name Duurdrift Sud. 119 poles to foliation surfaces, 4 poles to fold axial planes 2 fold axes, 43 lineations.

parageneses defining the two foliations are different in metamorphic grade. Such variable contact relationships illustrate that care must taken in interpreting 'regional foliations'. If only refoliated contacts were seen in the course of fieldwork, for example, this would fundamentally affect the relative age ascribed to the regional foliation in the surrounding gneisses. In discussing the eastern sub-area it was mentioned that the foliation along the entire eastern contact was conformable with that in the surrounding gneisses and it was suggested that the foliation in the whole sub-area could be the same as that seen in the granite. This conclusion is not compatible with the interpretation of the northern sub-area and it would not be logical to propose that the foliation in the northern and eastern sub-areas are separate entities. It is therefore proposed, following the more persuasive evidence along the Ham River, that the foliation between the granite and the East-Arm mylonite belt is not the same as that found in the remainder of the eastern sub-area.

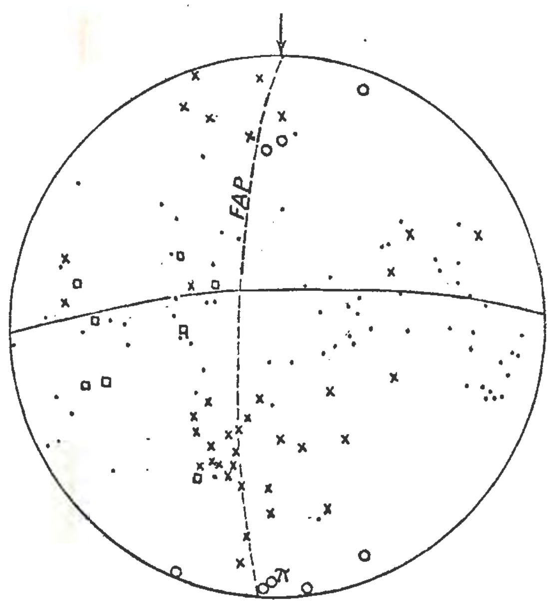


Figure 39. Orientation diagram for data collected from a 35 km^2 area south of the West-Arm mylonite belt and representing the Uheib antiform and its possible extension. 75 poles to foliation, 8 poles to fold axial planes, 8 fold axes, 37 lineations.

2. *Summary and aspects of the kinematic interpretation*

There is evidence for one set of macroscopic isoclinal folds in the northern sub-area. An older period of deformation is indicated by the foliation defining the folds and a linear fabric which is deformed around the large structures. A linear fabric is developed in the hinge zones of the large folds but not planar fabrics. Some refoliation of the layered gneisses is indicated in contact zones of the Naros granitoid which show a penetrative foliation common to both granite and host rocks.

The Naros complex is separated from the surrounding rocks by two mylonite belts - the West-Arm and East-Arm mylonite belts. Such zones are common around igneous intrusions and result from the different behaviour during deformation of layered rocks and intrusives which act as rigid incompressible bodies. During deformation high shear stresses and shear strains are developed at the contact of the rigid or more competent rocks (Strömgard 1973, p.232; Ramsay 1967, p.394). In layered rocks this produces features known as tectonic slides and bedding plane trusts whereas around intrusive bodies the same features are usually termed 'shear zones'.

Approaching the Naros granitoid from the north a marked change in the trend of the folds takes place from east-west to approximately north-south, roughly conformable to the outline of the intrusion. This change in trend is not due to any identifiable refolding but probably reflects the presence of the large intrusive mass which has controlled the structural pattern. Similar effects have been noted by Ramsay (1963b) around the external massifs of the Western Alps and by Nicholson (1965) around mantled gneiss somes in Uganda. Ramsay (1967, p.386) has suggested that all mantled gneiss domes are an expression of this same controlling influence.

E. WESTERN SUB-AREA

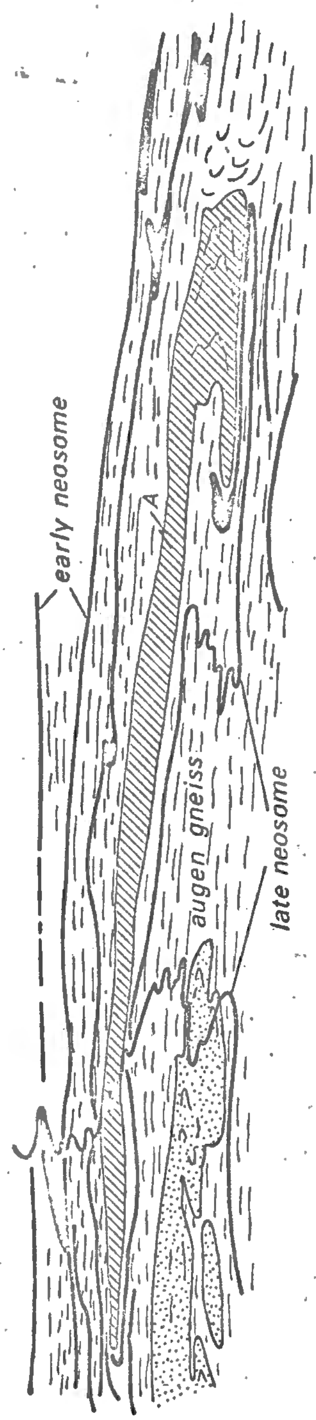
This is the largest of the three sub-areas and it also has the most complex structural framework. The geometry is dominated by a northwest trending shear zone (Pofadder ZAHNCAFS) of continental proportions. There is some evidence to suggest that this shear zone involved the production of several large northeast trending antiforms namely the Jackal, Ostrich, Mission and Falls antiforms and a major synform, the O'Connell synform (see Annexure 2). The shear zone and these large structures are best discussed separately and will be dealt with in the following chapter. In this section the structures pre-dating the development of the Pofadder ZAHNCAFS will be described.

Even though younger folds are clearly defined this does not significantly help to recognise older structures. In fact, the tightening and reorientation associated with the development of the ZAHNCAFS, especially in the southern part of the area, makes it extremely difficult to successfully trace older folds. The description of the older folds will therefore be confined to those folds and associated fabric elements which are visible in the northern part of the sub-area.

Nowhere is there evidence for macroscopic interference patterns resulting from the intersection of folds predating the late northeast trending ZAHNCAFS related folds. Examination of minor structures, however, show that not all of these early folds belong to the same generation. At least two sets of early macroscopic folds are indicated.

The largest of the older structures is the Leopard synform, an isoclinal northerly plunging fold featuring highly deformed rocks lying between the Falls and Mission antiforms (Annexure 2). On first inspection it might appear that this synform is a predictable fold structure linked with the bounding antiforms. The distinction in structural style, with highly deformed rocks in the synform and weakly deformed rocks in the antiforms, is not sufficient evidence in itself to contradict this interpretation (Ramsay 1967, pp. 382-384; Coward 1973b). The difference in age can be demonstrated quite clearly by the fabric elements, especially the axial plane foliation in the synform which is folded around the nose of the Falls antiform.

The Leopard synform is unique in the Onseepkans area since it is the only macroscopic fold with an extensively developed axial plane foliation. The fold is also notable for the vast amount of neosome material found in its hinge zone. The relationship between neosome, meososcopic folds and axial plane



foliation is well illustrated in Fig. 40 which represents an outcrop on the farm Velloorsdrift. 'Intrafolial folds' - folds with thickened hinge zones and discontinuous limbs - are defined by variations in lithology within the augen gneisses. In some of the horizons forming the folds, foliation is seen parallel to the contact but this is entirely subordinate to the foliation in the augen gneisses which is nearly everywhere parallel to the fold axial planes. At least two crosscutting sets of neosomes are present in this outcrop, while elsewhere cross-cutting neosomes of up to five distinct ages have been recorded. The fold axial planes and fold axes defined by both generations of neosome are related to the folding of the layering, and both neosome and layering share a common axial plane cleavage produced by the alignment of augen. None of the folds refold any other fold or deform the axial plane cleavage. The older neosome bands show isoclinal folding similar to those defined by the layering while the younger neosome bands show folds with less acute dihedral angles.

There is a very strong relationship between the lineations and the axis of the Leopard synform. Rods formed by isolated neosome fold hinges (intrafolial folds), mullions and boudins are very common and these are accompanied by a pronounced mineral lineation. There is no evidence that this synform folds an early linear fabric, although a traverse across the synform in the vicinity of the Rooimond River shows a great diversity in the orientation of the lineation which is due to the obvious refolding effects of the Pofadder ZAHNCAFS fold structures.

Due north of Onseepkans, along the Velloor River Valley, the intrafolial and isoclinal folds described above are seen to be refolded by mesoscopic tight-to-isoclinal folds associated with the Bushman Hills antiform (Plate 30) producing type-I interference patterns (Fig. 25). This structure therefore post-dates the Leopard synform and provides evidence for a second episode of folding in the western sub-area.

This second generation of mesoscopic folds varies from tight-to-isoclinal. A linear fabric is present parallel to the fold axes but seems to be confined to

Figure 40. Typical effects of deformation in the core of the Leopard synform. The stippled and ruled horizons are biotite gneisses of differing composition. The area of dashed lines represents augen gneiss. Heavy black lines are migmatite neosomes.

the hinge zones of the folds. The folds in Plate 30 show a variation in fabric style between synforms and antiforms; in the latter a penetrative axial plane foliation is developed while in the former linear fabrics are better displayed. This may be due to some geometric control but most probably it reflects the local variation in lithology - the antiforms are composed of a more uniform rock type than the synforms at this outcrop. Just north of the Bushman Hills antiform axial trace the two generations of folding are well developed. The younger generation parasitic folds are assymetric and overturned, facing east with a long flat lying limb and a short steeply dipping limb. This results in the majority of the older structures having approximately the same orientation as the younger folds and where refolding effects are not apparent it is exceedingly difficult to distinguish between the two sets (Fig. 41).

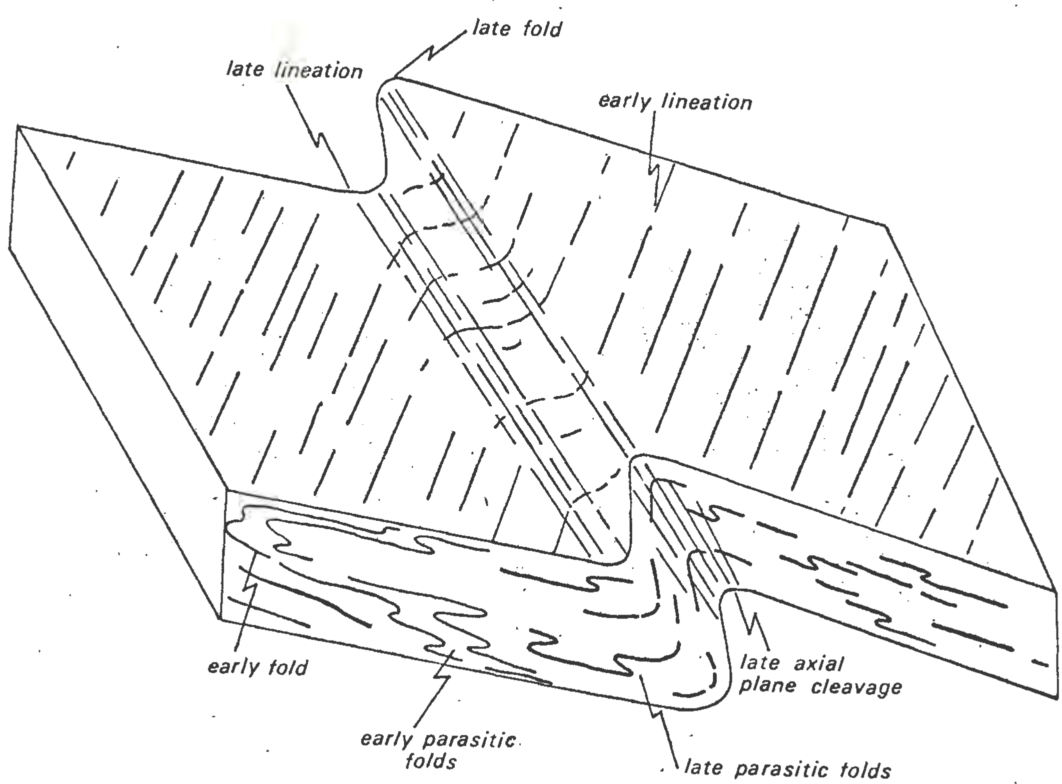


Figure 41. Assymetric mesoscopic late folds along the Velloor River valley forming type I interference patterns (Fig. 25) with earlier structures. Where refolding effects are not obvious it is difficult to distinguish between early and late minor parasitic folds.

The interference effects between the two sets of mesoscopic isoclinal folds can be traced across the Ostrich antiform. On the western flank of this structure identical refolding effects are visible but plunges are invariably towards the north, thus showing that the antiform is unrelated to these mesoscopic folds.

The planar fabric associated with some of the second generation mesoscopic folds is not seen in the Nautsis granite which forms the core of the refolded Bushman Hills antiform. The foliation here is clearly deformed by the macroscopic fold. It is almost certain that the Bushman Hills antiform and the Fortress antiform represent parts of a single fold made discontinuous through later deformation (c.f. Chapter V). There is no evidence from exposures along the Velloor River that migmatisation accompanied the second generation of folding.

Several macroscopic early folds are present in the western sub-area but it is not always certain to which generation they belong. The features shown by the older generation (exemplified by the Leopard synform) are:

(i) large amounts of migmatite neosome material formed contemporaneously with the folding.

(ii) very strong linear fabric indicated by mineral alignment, boudin alignment, intrafolial fold hinges, ribbing and sometimes surface intersection.

(iii) penetrative axial plane foliation visible in some parts of the major structures.

(iv) probable coaxial folding with the late megastructures formed in the Pofadder ZAHNCAFS. This is indicated by the linear fabric in the Leopard synform showing only a slight reorientation around the nose of the Falls antiform.

In contrast the younger folds (exemplified by the Bushman Hills antiform) show no associated migmatisation, weak planar and linear fabrics, and a type-H (Fig. 25) interference pattern with the Pofadder ZAHNCAFS megastructures. There is also evidence for the deformation of fabric elements by the younger fold set not seen in the earlier generation.

The Shaw synform on Keimas Farm probably represents an older structure. A strong linear fabric and contemporaneous migmatisation is visible in the hinge of this fold and a slight but persistent change in orientation of the linear fabric from north to northeast is visible as the result of the refolding of the lineation about the Mission antiform. These observations fulfill requirements (i), (ii) and (iv). The Eland antiform on Pelgrimsrust and Orangefall Farms shows features consistent with points (i) and (ii) but the re orientation and pronounced tightening of this structure in the core of the Pofadder ZAHNCAFS makes positive identification uncertain.

In the eastern part of the sub-area two folds, the Lost Dog and Uitlander synforms, have been identified as probable second-generation structures. Neither fold shows any associated migmatisation or any linear fabric parallel to the fold axis. There is evidence for reorientation of lineations across the axial trace of the Uitlander synform and these lineations include neosome intrafolial fold hinges that are a feature of the Leopard synform. Data from the hinge zone of the Uitlander synform are presented in Fig. 42 which represents an area from the Washaway antiform (Pofadder ZAHNCAFS structure) axial trace to the contact with the Naros granitoid. Even in this relatively small area only a very low symmetry pattern is achieved, possibly indicating that some fluctuation is due to earlier, unrecognised, structures. In the extreme west of the sub-area a northward plunging synform has been identified which shows none of the features associated with the older folds but cannot unequivocably be identified as a second-generation structure. It may be that this fold is the synform associated with the Jackal antiform. Exposures along a small river valley traversing the western limb of this fold (just outside the mapped area) shows a strong northwest plunging linear fabric. The eastern limb and hinge zone contain more variably oriented lineations and it would appear that the fabric is not associated with the fold.

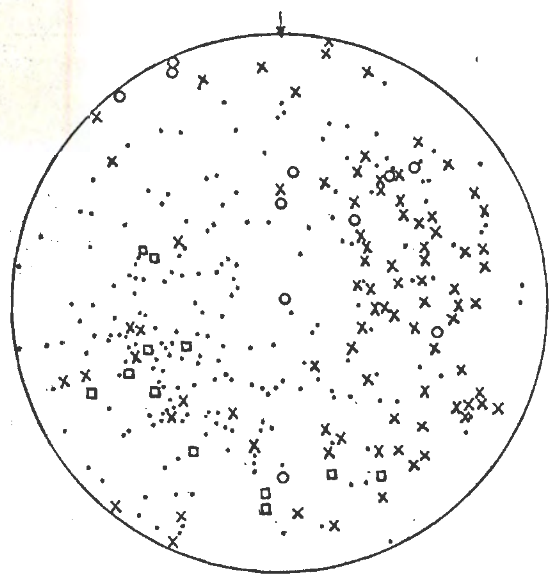


Figure 42. Orientation diagram of data collected from an 80 km^2 area between the Washaway antiform and the Naros granitoid. 184 poles to foliation surfaces, 12 poles to fold axial planes, 10 fold axes, 89 lineations.

1. *Planar fabrics*

It was seen in the northern sub-area that a variation existed in the contact relationships of the Naros granitoid with the intruded rocks which indicated that at least two foliations were present. A similar case can be made for the western sub-area but here the contact relationships of the granitoid show that a later, post-Naros, foliation is dominant and penetrative through country rock and granitoid alike. This may simply be a reflection of the increased scale of migmatisation in this sub-area since, in those localities where the granitoid appears to intrude the country rocks, the latter are only weakly migmatised or non-migmatised. Plates 18a, b, c showing later re-foliation are all taken from this sub-area.

In the Leopard synform it was seen (Fig. 40) that a refoliation of the gneisses accompanied migmatisation but that an earlier foliation is discernable; it is this fabric that probably represents the 'regional foliation' in the sub-area. The foliation of the Naros granitoid, where it is in contact with the migmatites, would not imply then that this is the fabric penetrative over the entire sequence in the sub-area.

2. *Aspects of the kinematic interpretation in the western sub-area*

It was mentioned above that the Leopard synform contains abundant neosome bands of more than one generation. Where younger neosome bands cross folds in the layering they do not appear to reflect the flow pattern in the rock. In rare cases the folded neosome bands may have an opposite sense of movement, i.e. where the neosome crosses an antiformal fold core it forms a synformal fold (Fig. 40). These features can only be explained if it is assumed that the folds in the layering were no longer active during the last stages of deformation. The rocks may be envisaged as undergoing a 'post-fold' flattening (Ramsay 1962, 1967, pp. 411-415; Hudleston 1973a, pp. 35-43) in which the isoclinal folds in the layering are deforming homogeneously, whereas younger neosome bands were still deformed by buckling. Both old and young folds have the same orientation showing that their development in relation to the finite strain ellipsoid is identical.

Only in special environments may it be demonstrated that the augen did not form *in situ* during this phase of deformation. If a folded competent horizon (A in Fig. 40) is examined it can be seen that on the inner arc of the fold the augen are exactly parallel to the axial plane whereas on the outer arc there is a small area where the foliation defined by the augen alignment is actually folded with the competent horizon and only away from the influence of this horizon do the augen again define a fold axial plane (Plate 31, hinge area of horizon A, Fig. 40). This can be explained by assuming that the competent horizon deformed by tangential-longitudinal strain (Ramsay 1967, pp.397-403; Ramberg 1963a). On the outer arc of this fold in the less competent material there exists a zone of low-contact strain which is essentially a zone of extension where earlier fabrics in the rocks stand the best chance of preservation (Ramsay 1967, pp.416-417; Ramberg 1961a).

The best examples of augen defining axial plane cleavage are found in the augen gneiss horizon. A much weaker axial plane foliation is found in the hinge of the Leopard synform south of the SII norite body. It is unfortunate that the exact relationship of this horizon to the axial trace of the synform is not accurately known north of SII. Only an inferred axial trace can be shown here since there is no change in orientation in foliation across the isoclinal fold. It may be that the mechanism of the formation of the Leopard synform is through tangential-longitudinal strain as indicated by the mesoscopic folds but it appears more likely that the surrounding quartzo-feldspathic gneisses have behaved more competently than the biotite gneisses in the synform (Coward 1973a, p. 144) in which case the synform probably developed by flexural flow folding.

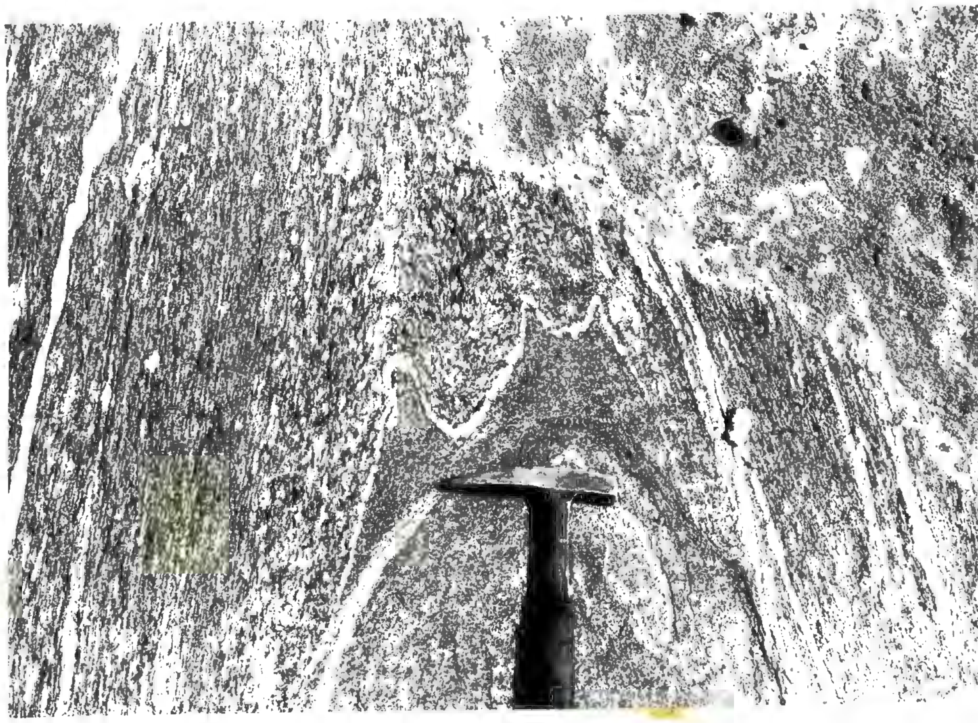


Plate 31. Hinge zone of the fold defined by horizon A. Fig. 45 showing that the augen are folded in a restricted zone at the outer arc of this fold. Velloorsdrift Farm.

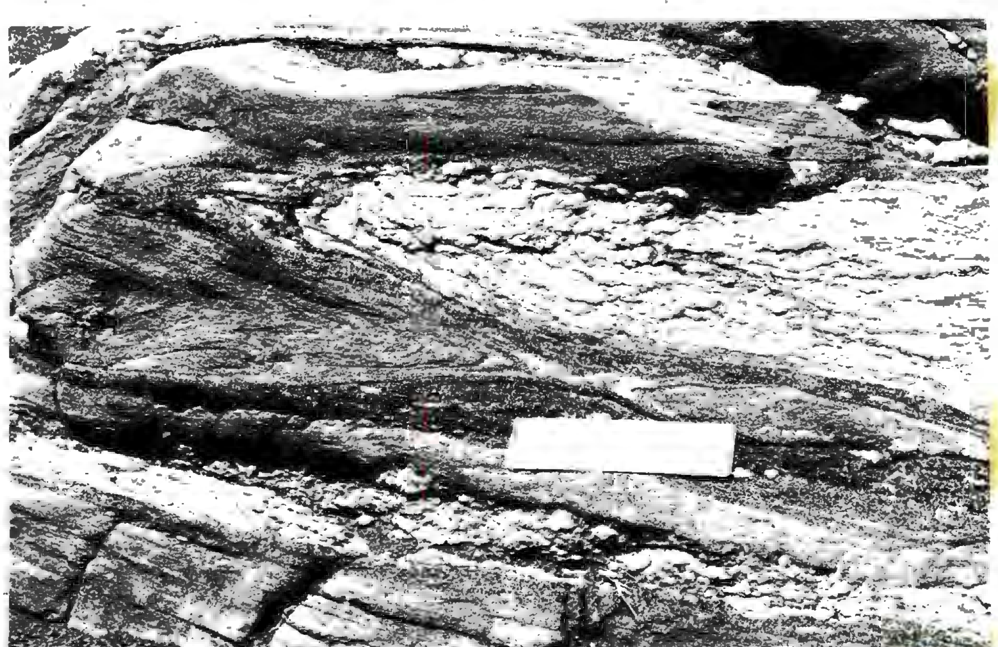


Plate 32. Dome-shaped fold representing a culmination point along the fold hinge or the result of variations in compressive strain during fold formation. Keimasmond Farm.

A good indication of the deforming mechanism would be provided by a comparison of the refoliation in the augen gneisses on the limbs and in the hinge of the fold, but as noted, the position of the augen gneiss horizon in relation to the fold axial trace is not known and in the hinge of the fold the horizon is not present. Ramsay (1967, pp. 391-397) has shown that in flexural flow folds the magnitude of the shear strain and the shortening in the rock is greater on the fold limbs than in the hinge zones and the finite strain is directly related to the dip of the folded horizon (Fig. 43a). The relationship between shear strain and total shortening in the rock will be discussed more fully in the following chapter but it can be seen that the intense refoliation and strong isoclinal folding in the augen gneiss horizon may be due to its presence on the limb of the Leopard synform (Fig. 43b).

It can also be suggested that this variability is possibly due to the finite strain state which reflects the superposition of a younger (Leopard synform) strain on an older (pre-Leopard synform) strain state (Fig. 44). If the Y and Z axes of the ellipsoids are interchanged the strain pattern moves into the constriction field and a linear fabric will be produced (Flinn 1962). Where these ellipsoids share mutually coincident strain axes a planar fabric will be preserved. The apparent dominant role played by linear fabrics in the Leopard synform implies that the deformation path B (Fig. 44) is probably that followed by the majority of the rocks during the younger deformation.

The Leopard synform is also notable for the presence of folded boudins (Plate 1). According to Talbot (1970, pp.55-56) these are significant in that they cannot be explained by progressive pure shear deformation. In such situations planes originally shortened may be extended since there is a possibility that they may pass from a shortening to an extension field (Flinn 1962, pp. 403-404). Folded boudins, however, imply rotational strain since it is only in this type of deformation that planes may pass through the surface of no-finite-longitudinal strain from an extension field to a shortening field. What cannot be determined in the Leopard synform is whether these effects are due to a later phase of deformation affecting the rocks.

Structures known as intrafolial folds (Turner and Weiss 1963), rootless folds (Rast 1956) or tectonic fish (McIntyre 1951) are common in the Leopard synform and these structures are widely accepted as indicating a complex deformational history (Whitten 1966, p. 191 and Fig. 316). Although this interpretation may be correct, and the intrafolial folds are older structures, many authors do not appear to have appreciated that such folds are a normal product of progressive deformation. During pure shear deformation lines and planes in a rock will rotate towards the XY plane of the strain ellipsoid. However, material planes rotate faster than imaginary surfaces and for certain initial orientations material planes will migrate through the surface of no-finite-longitudinal strain (Flinn 1962; Ramsay 1967, p.117), the surface undergoing compression therefore passing into a field of extension. This may result in a folded surface unfolding or necking, in which case an 'intrafolial fold' will form (Fig. 45).

An alternative and possibly more general case for the formation of intrafolial folds during a single progressive deformation will be suggested in Chapter VI.

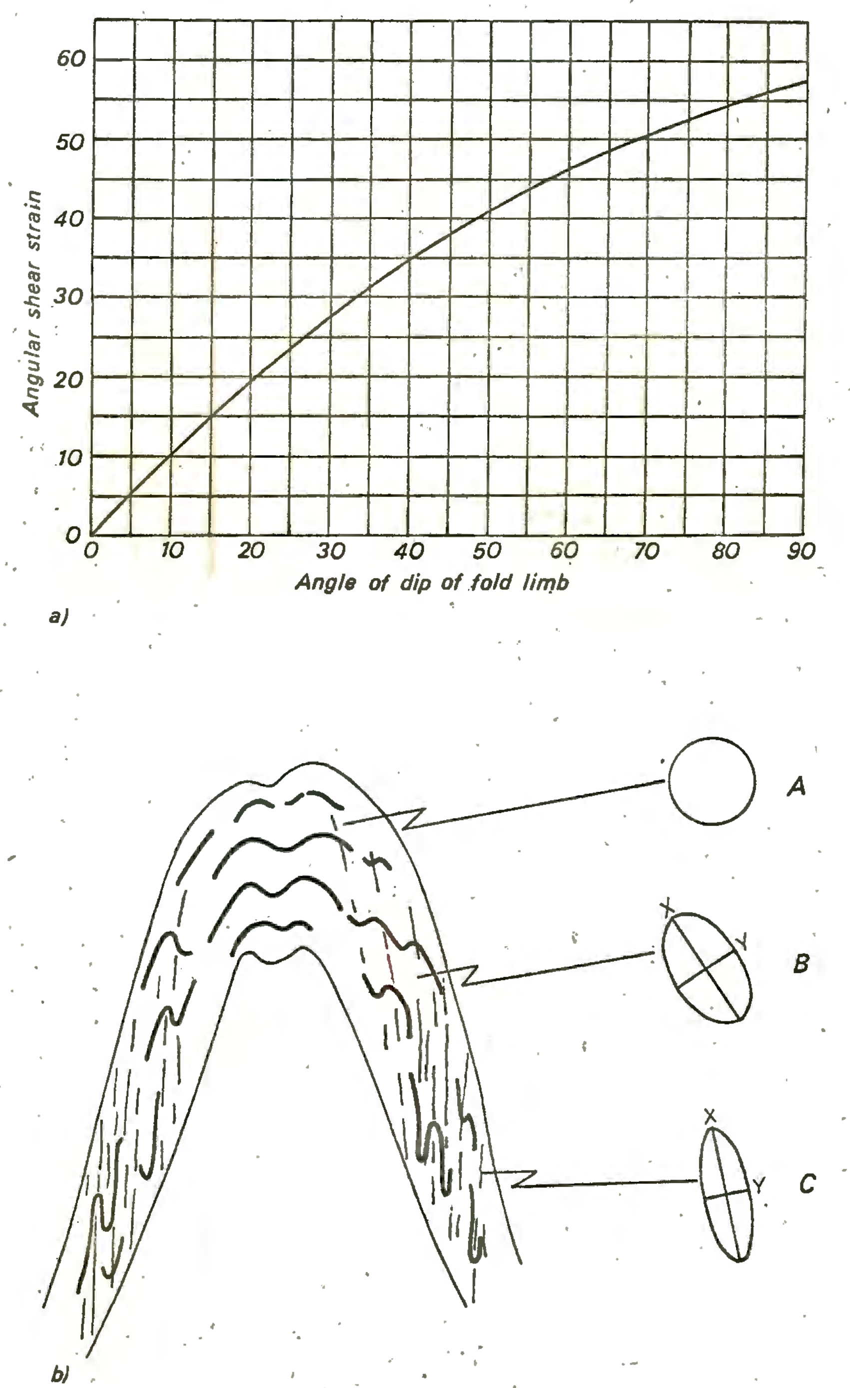


Figure 43. (a) Variation on shear strain related to the dip of fold limbs
(b) Variation in finite strain and associated fabric elements in a fold formed by flexural flow. After Ramsay (1967, Fig. 7-57).

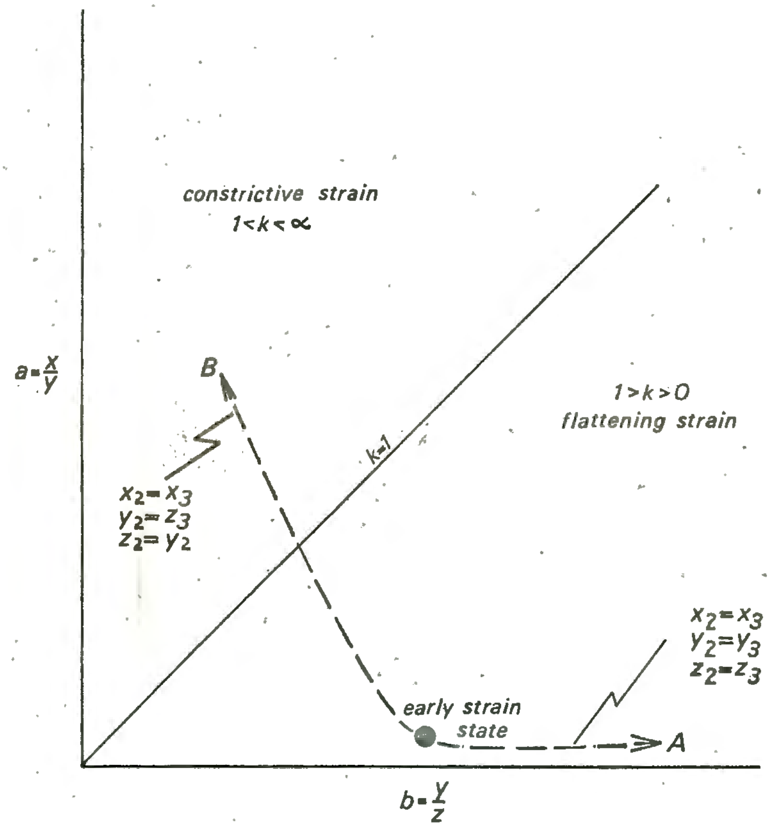


Figure 44. Finite strain explanation for variability of the Leopard synform tectonic fabric. Where earlier and later ellipsoid axes maintain a constant orientation a planar fabric is produced. (A) where Y and Z axes are interchanged (B) a linear fabric results. Dashed lines indicate the possible deformation paths from an early strain ellipsoid of the type 1 > k > 0 or k = 0. After Coward (1973b).

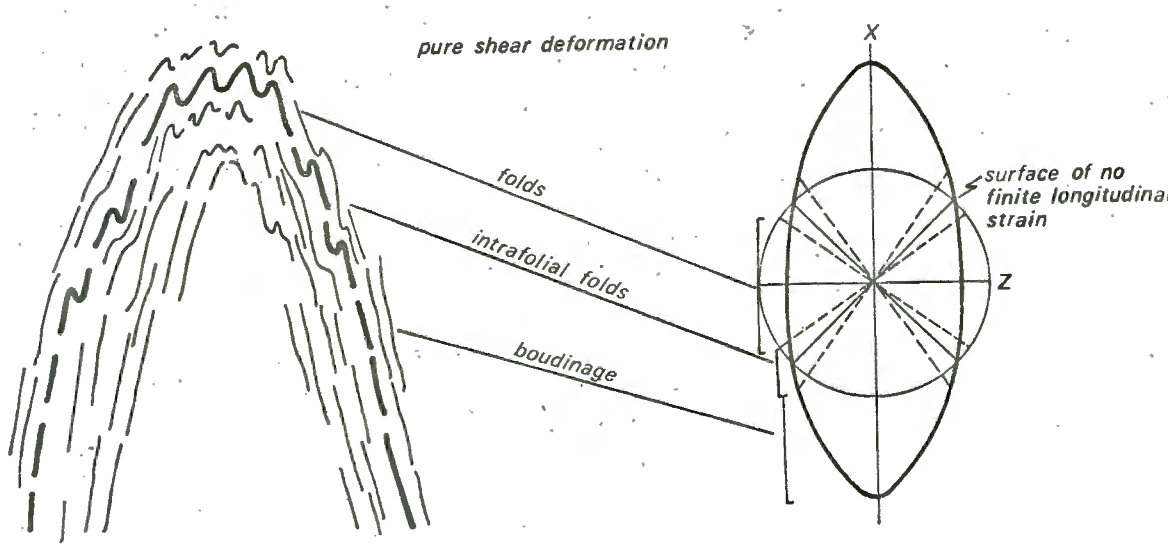


Figure 45. Formation of intrafolial folds through progressive pure shear deformation. After Ramsay (1967, Figs. 3, 62 and 3, 64).

Small dome-like structures or eye folds are occasionally encountered in the Leopard synform (Plate 32). These possibly result from fold interference but similar features in the Alps have been related to the variation in compressive strain during a single fold forming event (Chadwick 1968). Borradaile (1972) has used an argument based on strain ellipsoid considerations to explain these folds (Fig. 46). He has pointed out that if folds form in a constrictive strain environment where $\infty>k>1$ (Flinn 1962) the fold axis must change its orientation during progressive deformation. If the fold axis is parallel to the Y direction of the ellipsoid (Fig. 46a) it will start producing a series of culminations and depressions (Fig. 46b). Since the XY plane of the strain ellipsoid containing the fold axis intersects the surface of no-finite-longitudinal strain the fold axis could pass through the surface and suffer extension, possibly resulting in a finite strain state with nearly rectilinear fold axes once more (Fig. 46c).

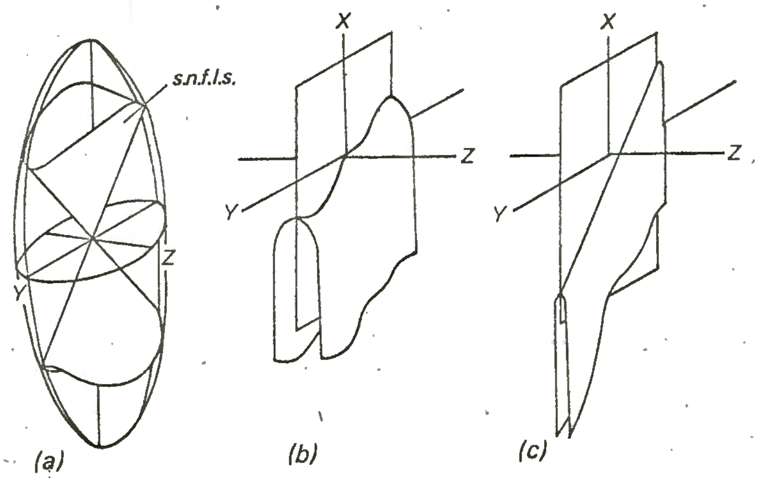


Figure 46. Development of a fold axis in a constrictive strain environment where the fold axis is parallel to the Y direction of the strain ellipsoid. (a) Form of the strain ellipsoid showing the intersection of the surface of no-finite-longitudinal strain (snfls) with the XY plane of the ellipsoid. (b) Initial formation of culminations and depressions in the fold axis (c) Possible formation of rectilinear fold axis as it migrates through the snfls. After Borradaile (1972).

Throughout most of the sub-area the Beenbreek granite shares a foliation with the intruded gneisses suggesting that pre-Beenbreek (i.e. Kumian) planar fabrics as seen in xenoliths must have been destroyed. The development of foliation in the granite is well illustrated by an outcrop on Keimas Farm, reproduced in Fig. 47. Foliated xenoliths in the granite are up to 1 m in diameter but within a distance of 2 m they are transformed into thin discontinuous layers, some 15 cm thick. Such strong flattening strains have been envisaged by Flinn (1962, p.388) as being due to the k factor of the strain ellipsoid being of the type $1>k>0$. This classification of tectonic fabrics does

Figure 47. The penetrative foliation of the Beenbreek granite containing xenoliths. A. angular xenoliths, some foliated. B. strongly foliated contact between granite and augen gneiss. C. original intrusive contact. D. augen gneiss resulting from the foliation of Beenbreek megacrystic granite. E. strongly flattened xenoliths lying parallel to the foliation. F. migmatite leucosome bands. G. country rock (Jerusalem formation). H. undeformed Beenbreek granite. Outcrop on Keimas Farm.

not take into account possible volume reductions (Ramsay and Wood 1973) and therefore the fabric should be described as one of apparent flattening.

The concept of a strong flattening strain having affected these rocks may help to explain a further aspect of their character. Many of the norite and amphibolite bodies in the Beenbreek granite have lenticular shapes which, on first examination, suggest that the former are intruded into the granite as dykes. It may have been on such evidence that Beukes (1973) mistakenly interpreted the age relationships of his Eendoorn granite (equivalent of the Beenbreek granite) and the SI norite body. On the farm Beenbreek, dyke-like bodies of amphibolite and norite are very common but if these are traced from high-strain areas (foliated granite) into unstrained or low-strain zones it is always found that unambiguous contact relationships reflect a younger age for the granite. In the writer's experience a foliation in the megacrystic granite, however slight, indicates a high flattening strain that is otherwise not suspended. Contact relationships with other rock types should therefore be interpreted accordingly.

Some of the larger norite bodies (especially SII) are ramified by a network of planar, foliated, leucocratic horizons that often contain isoclinal folds within the confines of the horizon (see also Joubert 1971, p.61). The norites surrounding these horizons is usually non-foliated. The explanation for these deformed gneissic horizons lies in the behaviour of the norites during deformation. Acting as large rigid bodies, strain is highly localised along pre-existing anisotropies - in this area pegmatites, and deformation results in the foliation of the pegmatites while the adjacent rock is unaffected. The termination of these horizons shows how thin veins of pegmatite have become boudinaged and folded (Plate 33 and Fig. 48) with a foliation penetrative through both the folded veins and the amphibolitised norite.

F. SUMMARY

Having dealt with the three subdivisions of the Onseepkans area it is now possible to discuss the correlation of the structures and fabric elements seen in them. It has been shown that, within each sub-area, there is considerable variation in the orientation of the major structures and both the spatial distribution and intensity of the fabric elements associated with them. This variability extends to outcrop scale where related mesoscopic folds show diverse fabric elements. Although it has been established that planar fabrics are present representing the M_2 and M_3 metamorphic episodes the information cannot be effectively utilised for correlation purposes since both fabrics are defined by mineral parageneses with a similar metamorphic grade (c.f. section IIIE). These considerations seriously handicap attempts to present a unified structural interpretation and, at best, a highly subjective picture will be obtained. It is emphasised that the deformational history presented here is only one alternative of many that could affect the data, however, it does have the advantage of being the simplest interpretation.

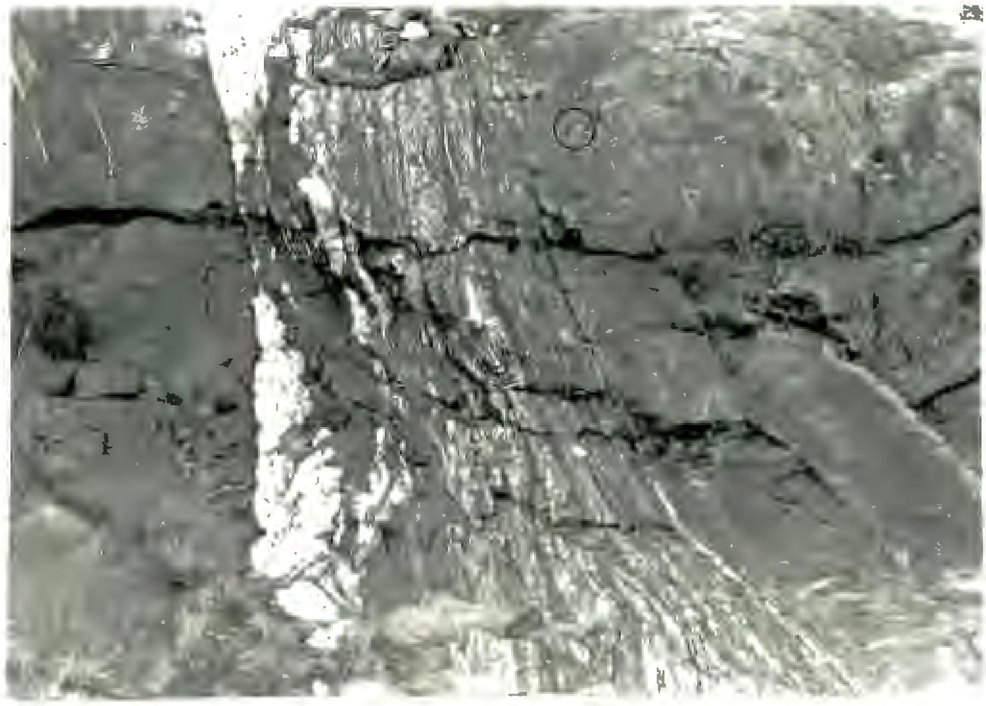


Plate 33. Termination of a leucocratic horizon in norite. The individual veins are strongly folded and boudinaged with a foliation in the norite axial planar on the folds. The circled area represents Fig. 48. Keimasmond Farm.

Plate 34. Typical alternation of gneiss and mylonite in the Pofadder Lineament. Pelladrift Farm.

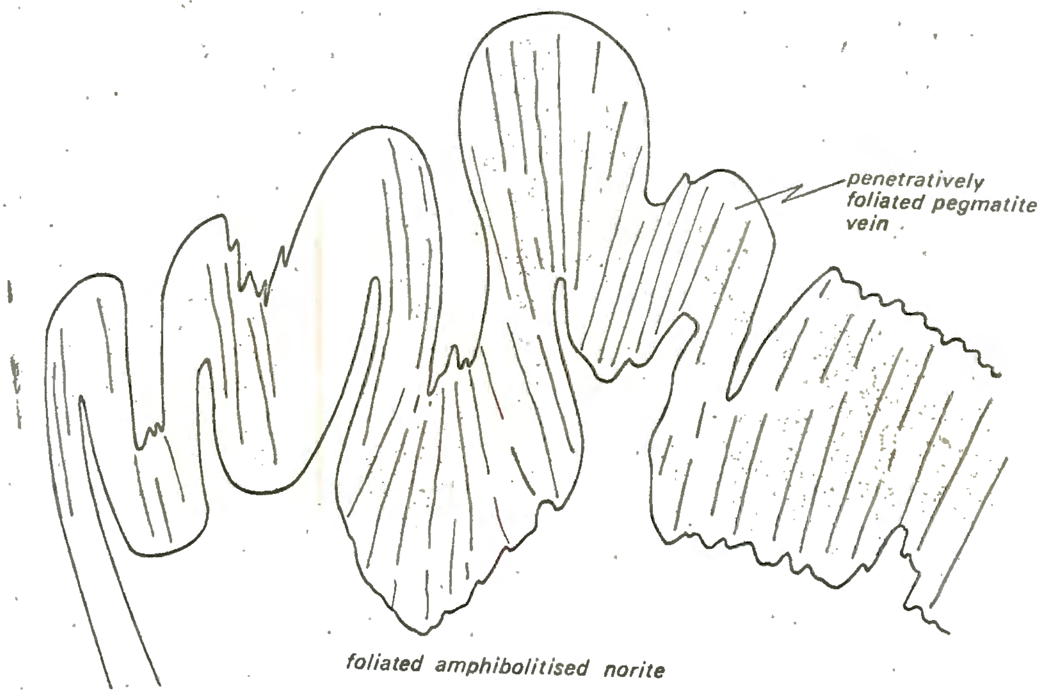


Figure 48. The deformation and foliation of a narrow pegmatite vein in foliated amphibolitised norite. This figure represents the circled area in Plate 33 (natural scale).

The earliest deformational events to have affected the rocks in the Onseepkans area are largely conjectural. A pre-Beenbreek-granite tectonic fabric in granulite-grade metamorphic rocks has been identified in xenoliths. The paucity of granulite-grade assemblages now present in the area and the penetrative foliation of the Beenbreek granite implies that fabric elements associated with the earlier events have largely been obliterated. It has not been possible to relate this early fabric to any fold. The deformational episode represented by this fabric is referred to as D_1 and the foliation as s_1. The event accompanied the M_1 metamorphic episode.

Following the intrusion of the Beenbreek granite, a period of deformation resulted in the development of a penetrative planar fabric. A major problem exists over whether it is this fabric or a planar fabric post-dating the Naros granitoid that now represents the 'regional foliation' over most of the area. In the northern sub-area the Naros granitoid intrudes foliated rocks in some outcrops but in others shares a penetrative foliation with the surrounding gneisses. The problem cannot be solved at this scale of mapping but if one accepts that the post-Naros foliation is connected with migmatisation it must be subordinate to a pre-Naros planar fabric which shows no associated migmatisation.

The pre-Naros planar fabric will be denoted s_2 and the deformational event D_2. The structures at present controlling the geometry of the Onseepkans area cannot be related to D_2. A multitude of 'intrafolial folds' now found throughout the area may be D_2 folds but the ambiguity surrounding the interpretation of these structures makes them unreliable markers. There was some suggestion in the eastern sub-area that the older of two sets of folds deforming the foliation also folded a lineation, but generally linear fabrics have not been identified with this event.

In the eastern and western sub-areas two sets of macroscopic folds deform the regional foliation. The later set in both cases may have an associated B-lineation but both fold a strong linear fabric. The macroscopic folds in the northern sub-area have many features similar to the younger set of folds in the other sub-areas in that they deform a strong linear fabric, have B-lineations developed in their hinge zones and generally have no associated planar fabrics. In the western sub-area the older linear fabric is often represented by intrafolial fold hinges and boudins defined by neosome material. This is also the case for the older lineation in the northern sub-area and, to a lesser extent, in the eastern sub-area. Since it has been proposed that only one period of migmatisation is present, a relationship is suggested between the early macroscopic folds in the eastern and western sub-areas whereas the macroscopic folds in the northern sub-area may be related to the younger folds in the east and west. A structural history based on this interpretation is presented in Table 15; this should be compared with Table 12 in Chapter III. The relative ages of macroscopic folds shown in Annexure 2 have been derived from this table.

TABLE 15

Possible correlation of macroscopic folds and fabric elements between the three subdivisions of the Onseepkans area. Pofadder ZAHNCAFS folds excluded. Compare with Tables 12 and 13

Western sub-area	Northern sub-area	Eastern sub-area	Event
Pofadder ZAHNCAFS			D_5 + D_6
Macroscopic and mesoscopic folds (late folds) Local weak planar fabric s_4 Local B-linear fabric	Macroscopic and mesoscopic folds. No planar fabric B-linear fabric developed in hinge zones l_4	Macroscopic and mesoscopic folds (late folds) B lineation locally developed Weak planar fabric s_4	D_4
Macroscopic and mesoscopic folds (early folds) Locally penetrative planar fabric s_3 Strong B-linear fabric l_3 Contemporaneous migmatisation	Mesoscopic folds Locally penetrative planar fabric s_3 Strong B-linear fabric l_3	Macroscopic and mesoscopic folds (early folds) Strong B-linear fabric especially in hinge zones l_3 Weak planar fabric s_3	D_3
	Intrusion of Naros granitoid		
Foliation of Beenbreek granite and country rocks - probable 'regional foliation' s_2 Lineation l_2?			D_2
Intrusion of Beenbreek granite			
Foliation in xenoliths in Beenbreek granite s_1			D_1

CHAPTER V

ZONES OF ANOMALOUSLY-HIGH NON-COAXIALLY-ACCUMULATING FINITE STRAIN (ZAHNCAFS)

In recent years lineaments in the earth's crust have excited considerable attention following the development of plate tectonic theory. The study of lineaments in basement rocks has tended to lag behind and it has only been comparatively recently that their significance has been recognised (Nickelsen 1975). The problem was due in part to the rather unique character that these lineaments assumed at lower crustal levels and it was not until the publication by Ramsay and Graham (1970) of new methods for their investigation that any significant progress in understanding their formation was made.

If the specialized terminology of plate tectonic theory is avoided the names most generally used for lineaments are faults, fault zones, thrusts, shear fractures, fracture zones, mylonite belts and shear zones. The main criticism of earlier, and indeed fairly recent, publications using this terminology is that the authors have not appreciated that the above terms cannot be interchanged and, to a large extent, indicate mutually exclusive rheological conditions. The essential feature of faults, fractures or shear fractures (Griggs and Handin 1960) is that they occur in the elastic range or rocks where failure has occurred with no accompanying deformation (Fig. 49), i.e. they signify brittle behaviour and are found in the upper zone of the crust where rocks behave as brittle or semi-brittle solids (Price 1966, p.57). Shear zones and most mylonites, on the other hand, can be described as planar zones where deformation has occurred *without* loss of cohesion in the rock i.e. they indicate ductile or 'plastic' behaviour of rocks (Fig. 49), and as such represent deformation at lower crustal levels (Ramsay and Graham 1970, p.800).

In many standard texts the authors omit to make a distinction between faults and shear zones or, if recognised, simply fail to describe the latter (Ragan 1973, p.157). In some texts features which are clearly related to ductile deformation are described as faults (Hills 1963, Fig. VII-9; Spencer 1969, p.70; Billings 1972, p.202 and Fig. 7-24). De Sitter (1956, p.118) and Denis (1972) are among the few authors of a standard text to have pointed out the implications of ductile and brittle behaviour in relation to the development of a rock fabric but Ramsay (1967) is the only author to have

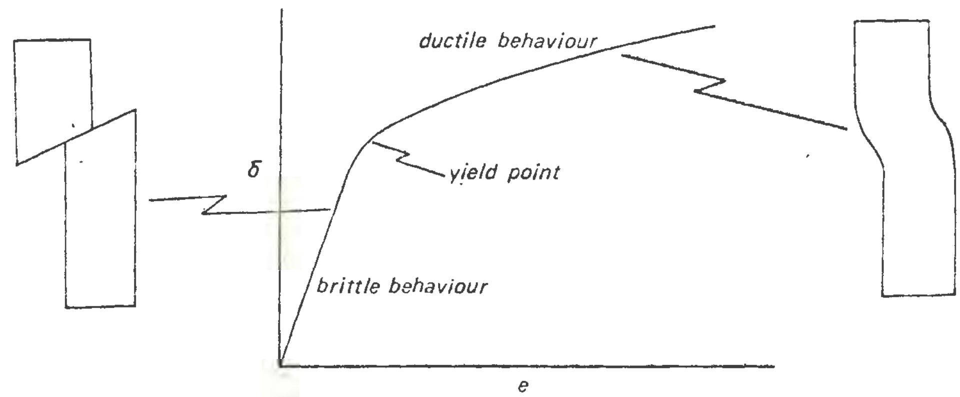


Figure 49. Generalised stress-strain curve showing the distinction between brittle and ductile behaviour.

described these implications in some detail.

A contributory factor involved in this problem is the confusion caused by similar sounding terms *viz* shear (as in pure shear or simple shear), shear zones and shear fractures. The word shear has very little real meaning in structural geology since all deformed rocks are sheared rocks by definition (i.e. shear strained). The common description of rocks in the geological literature as 'sheared' is therefore misleading and most often, but by no means always, appears to stand for 'shear fractured'. In this sense Tchalenko (1970) uses the term 'shear zone' in quotation marks to describe an area of shear fracturing. A shear zone as defined by Ramsay and Graham (1970) and Hobbs (1972) is a zone in which deformation has taken place by simple shear (see also Ramsay 1967, pp.83-91). The term simple shear refers only to a special type of strain pattern which is distinguished from pure shear because particles in a rock are conceived as being displaced along a series of parallel planes - the x direction in Fig. 50A (Jaeger and Cook 1969, pp.425-431). Shear fractures of faults are planes of high resolved shear stress (Fig. 50B) where shear stress denotes the stress acting on any plane which is not perpendicular to the normal stress (Jaeger and Cook op. cit., pp. 9-24). Thus a clear distinction must be made between simple shear and shear fracturing. The former is a kinematic concept and the latter is a dynamic concept and the word 'shear' which they both have in common has entirely different implications.

An important point to emerge from Ramsay and Graham's (1970) paper was that the change in orientation of fabric across these lineaments is solely the result of heterogenous simple shear and therefore the entire zone which encompasses this reorientation is an integral part of the shear zone (Fig. 51). They have also established the order in which the fabric is produced and they have proved that the core of the zone, where the mylonites are parallel or sub-parallel to the shear direction, represent a final stage in the development of a shear zone and they do not form in an initial stage as formerly believed. Many workers have referred to the cataclastic rocks commonly found in the core of these zones as a fault, fracture or 'shear zone' and the surrounding displacement as 'drag' or 'drag folds' (Moody and Hill 1956) implying that the

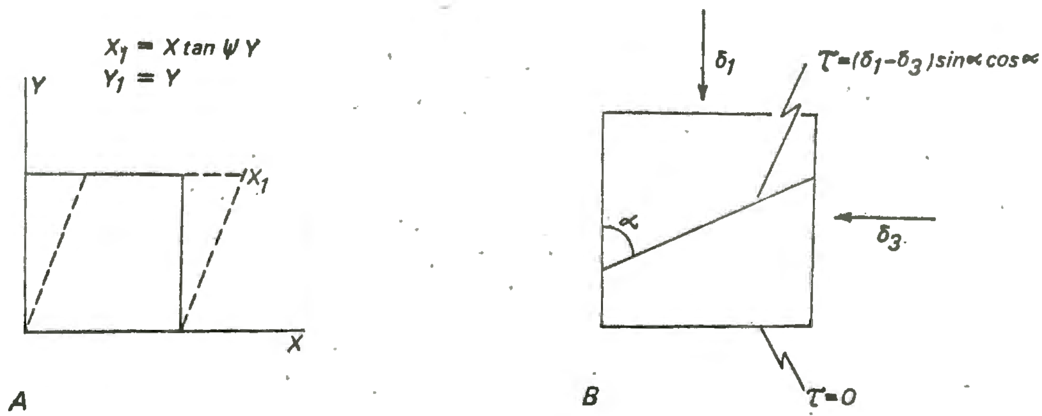


Figure 50. A. The strain pattern known as simple shear.
B. Normal and shear stresses acting on a unit square and on an internal surface making an angle α with the side of the square

pattern was the result of two deforming mechanisms. The word 'drag' is here used in quotation marks to indicate the geometrical effects usually implied by geologists. Drag is a perfectly legitimate concept but its application in geology is not well known. For example the shear zones described by Ramsay and Graham (op. cit.) may be drag-zones in the strict sense of the word.

De Sitter (1958b) and Ramberg (1963b) have pointed out the misconceptions involved in the use of the term 'drag fold' as applied to minor structures developed on fold limbs. Another common misapplication of the drag concept has been to the deformation seen at boudin terminations where friction exercised by surrounding rocks moving into the gap between separating boudins was thought to cause 'barrelling'. Gay and Jaeger (1975, p.328) have shown that barrelling is not due to drag but is the result of deformation in the rock *before* separation into boudins. The paper by Ramsay and Graham (op. cit.) may be regarded as one more objection to the concept of 'drag'. Indeed, it may now be asked if any *unequivocal* examples of 'drag' are known to exist (c.f. also Garfunkel 1966 and Dennis 1972, p.302).

Shear zone is, at present, the most common term applied to the type of deformation described by Ramsay and Graham (1970). However the writer feels that because of the ambiguity surrounding the word 'shear' a new term might be usefully introduced to overcome this difficulty. Unless shear zones happen to be superimposed on an area of deformed rock so that they cause the cancellation of the deformation the zones are usually recognised by their abnormally-high finite strain pattern. Simple shear is also a special type of deformation which is known as rotational strain because the principal strain axes do not remain fixed during progressive deformation (Ramsay 1967, p.83). It is therefore proposed that a name for these zones might be more revealing if it implied anomalously-high finite strain and rotational deformation. One further point to be considered is that raised by Elliott (1972) over the use of the term rotational strain. He has suggested (op. cit., p. 2627) that 'rotational strain' implicitly acknowledges the existence of 'irrotational strain', a concept that may often be refuted by the suitable choice of coordinate frames. In its place he proposed 'coaxially accumulating' in situations where the same material lines remain axes of all finite and strain rate ellipsoids and 'non-coaxially accumulating' where they do not. With this in mind a shear zone may be referred to as a Zone of Anomalously-High Non-Coaxially-Accumulating Finite Strain or ZAHNCAFS. The writer is fully aware that there is little likelihood of this term becoming accepted but feels that some attempt must be made to resolve the present ambiguity surrounding the use of the word shear.

One such ZAHNCAFS of continental proportions controls the geometry of the rocks in the western subarea and this chapter is largely concerned with its description.

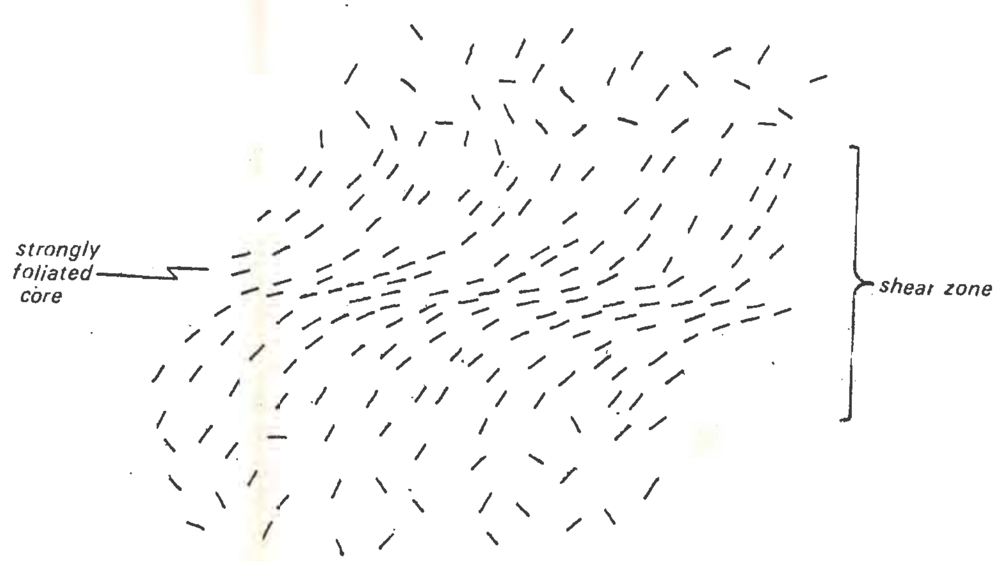


Figure 51. Development of a shear zone in isotropic rocks showing that the 'S' shaped pattern across the shear zone is the result of simple shear and is not caused by drag.

A. THE POFADDER ZAHNCAFS - DESCRIPTIVE

1. *The extent of the mylonite belt*

South of Onseepkans a northwest trending belt of mylonite can be traced from east of Pofadder to the Karoo cover rocks west of Warmbad (Fig. 24). These

These mylonites have been previously recorded as the Tantalite Valley Megaskuif-keurzone (Beukes 1973), Pilgrim Lineament (Toogood 1974), Tantalite Valley Mylonite Belt (Blignault et al. 1974) and Pofadder Lineament (Joubert 1974a). On a regional scale it seems best to adopt the term Pofadder Lineament since, as noted by Toogood (op. cit.) Pilgrim Lineament has only local significance. Beukes (1973) has shown this mylonite belt on his map as a single line which does not fully reflect its true width. At Tantalite Valley the belt is 7 km in width (Moore 1976) and in the west of the Warmbad area the writer found it to be 4 km wide. Where shown on Beukes' map at the junction of the two study areas the belt corresponds to a minor bifurcation of the main zone lying to the north of the Lineament shown in this report (see accompanying geological map).

The lateral extent of the Lineament as shown on the map should be regarded as a minimum. Evidence for mylonitic fabrics can be found outside this zone - especially to the south where small discontinuous mylonite belts a few metres in width are very common. The Lineament is not entirely composed of refoliated rocks but horizons of mortar gneiss, flaser gneiss and mylonite occur sand-wiched between unaffected gneisses (Plate 34) often with knife-sharp contacts. Discontinuous lenses of rock types representing several formations are tec-tonically interfingered in the Lineament and the writer considers it more im-portant to show these and the mylonites as one entity on the map rather than attempting to show them individually. It is not always apparent in the field that mylonite fabrics are present since slightly mylonitised fabrics are only visible in thin section.

West of Warmbad the exposures are poor but the extension of the Pofadder Lineament is shown by Beukes (1973) and Blignault et al. (1974) to be a con-tinuation of the same northwesterly trend as seen on the ERTS photo in the On-seepkans area. According to the geological map (Beukes op. cit.) the Linea-ment disappears under the Karoo cover rocks approximately at the homestead on the farm Harib. The writer visited this locality but was unable to find any trace of mylonitic rocks. Subsequent remapping of the area (Toogood 1975a) showed that the Lineament does not strike northwest from Tantalite Valley but swings west through the homestead on the farm Bankwasser and continues to the junction of the farms Kromrivier, Witputs, Soekwater and Sperlingsputs (see Fig. 52). In this locality it has apparently been confused with the contact between the Vioolsdrif complex and the Namaqualand gneiss terrain and is termed by Blignault et al. (1974, p. 32) as the Southern Front of the Namaqua Province. However, a map of this vicinity (Blignault in preparation) clearly shows a pro-gressive change in the orientation of this mylonite foliation from east-west at the junction at the four farms mentioned above to the overall regional northwest trend of the belt as the Karoo cover is approached.

It can also be seen from Fig. 52a that the extension of the 'Front' from this locality is supposedly towards the southeast and an area of mixed gneiss is shown (Blignault et al. Annexure 1) between this 'Front' and the Pofadder Lineament as now mapped. Traverses across this area made by the writer failed to reveal this continuation and no change was noted in the rock types which are predominantly granitoids of the Vioolsdrif suite. These granitoids show the non-penetrative development of mylonite fabrics that is a feature of the zone south of the Lineament at Onseepkans.

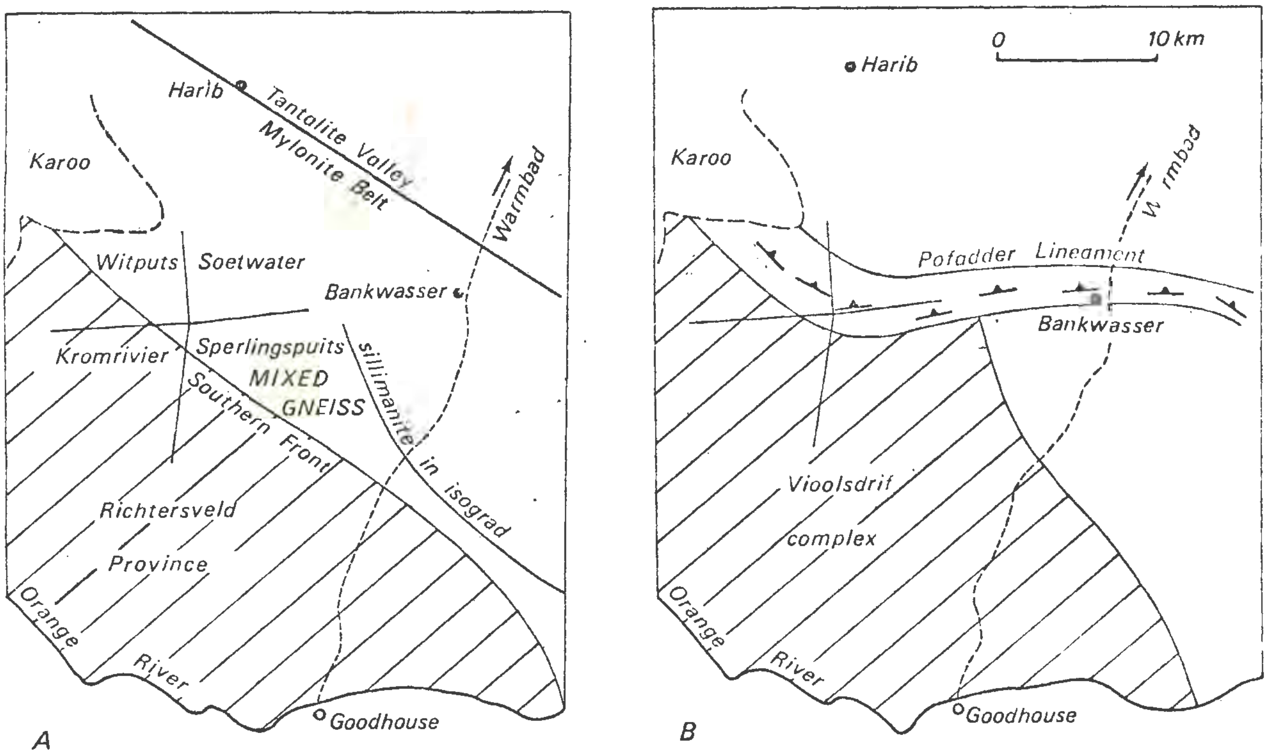


Figure 52. A. Map of the area north of Goodhouse according to Blignault et al. (1974) and Beukes (1973). B. According to Toogood (1975a).

Vioolsdrif granitoids are known to occur north of this 'Front' (Beukes 1973, Annexure 1) but according to Beukes (op. cit.) none occur north of the Pofadder Lineament as now mapped (Toogood 1975a). A contact could be placed between granitoid and gneisses exactly at the 'sillimanite-in' isograd depicted by Beukes (op cit. p.168). This isograd is shown by Beukes to extend across the Pofadder Lineament as now shown. This is rather puzzling as sillimanite is clearly shown on his metamorphic index map on both sides of the isograd north of the Lineament.

There appears to be fairly conclusive field evidence that the contact between the Vioolsdrif complex and the layered gneisses in the Warmbad area is represented by the Pofadder Lineament and is not one resulting from the progressive metamorphism of the Vioolsdrif complex as proposed by Blignault (1974, p.55). Continuations of the mylonite belt to the northwest are described by Blignault et al. (1974) and at Ai-Ais by Jackson (1976). A continuation to the southeast of the Onseepkans area is described by Joubert (1974a, 1974b). If this correlation is valid a mylonite belt of 450 km in known length is indicated.

2. *Mylonite fabric*

At present there appears to be no consensus on the terminology of mylonites (Spry 1969, pp. 227-231; Higgins 1971) and the same term is sometimes used with completely different definitions. Blastomylonite, for example, is commonly used for mylonite that shows evidence of recrystallisation (e.g. Johnson 1961) but Lundgren and Ebblin (1972) used the same term in exactly the opposite sense, that is for mylonitic rocks that do not show evidence of recrystallisation (op. cit., p. 2776). Mylonitic rocks appear to be rare examples of rocks that still rely on a genetic definition and there is a strong connotation of 'crushing', 'milling', 'granulation' and 'brecciation' when mylonites are described. This is unfortunate for recent work has shown that their origin of formation is far from clear (see section VB1).

In this report the term mylonite is used for rocks with large mineral grains showing evidence of strain and a matrix of smaller grains making up at least half the rock. Instead of using prefixes to denote varieties of mylonite separate terms will be used. A flaser gneiss (Katz 1968) is a rock in which large, unstrained mineral grains in a fine-grained matrix make up more than half of the rock. Rocks in which the mineral grains show only slight evidence of strain and the surrounding finer grained material forms only a thin selvage (mortar texture) are called mortar gneisses. No inferences about the temperature or mechanism of formation are implied in the use of these terms.

Mylonite fabrics have been well described in the literature (Christie 960, 1963; Johnson 1957, 1960, 1961; Katz 1968; Higgins 1971; Bell and ridge 1974) and none of the features associated with the Pofadder Lineament appear to be unique. The first evidence of mylonitisation is the ribbon-like development of quartz grains with a pronounced undulatory extinction and sometimes deformation lamellae (Plate 35). Mortar structure first makes its appearance here as thin incomplete selvages of fine-grained quartz around quartz porphyroblasts. Plagioclase is not strongly affected at this stage and tends to show only saussuritization features. The effects of mylonitisation on microcline, however, are immediately obvious and at crystal margins small exsolution lamellae can be seen extending into the crystal. This development of perthitic intergrowth is a well known feature of mylonitic rocks (Johnson 1961; Spry, 1969, p. 234; Chayes 1952). Biotite may show some evidence of alteration to chlorite but retrograde effects in other mafic minerals are absent. This stage of mylonitisation produced mortar gneisses but the effects described here are usually not visible in hand specimen.

As deformation increases the effects become sufficiently pronounced to be visible. Flaser gneisses representing this stage are recognised by porphyroclasts around which a foliation flows and swirls with trains of finer-grained material framing the porphyroclast. Chlorite can be seen in these rocks without the aid of a hand lens and when viewed from a distance the rock gives the impression of being more strongly foliated than the gneisses from which it was derived.

Under the microscope the quartz grains are seen to be strongly elongated and undulatory extinction is very pronounced with many crystals appearing to

Plate 35. Photomicrograph showing deformation lamellae in a quartz grain. X40.

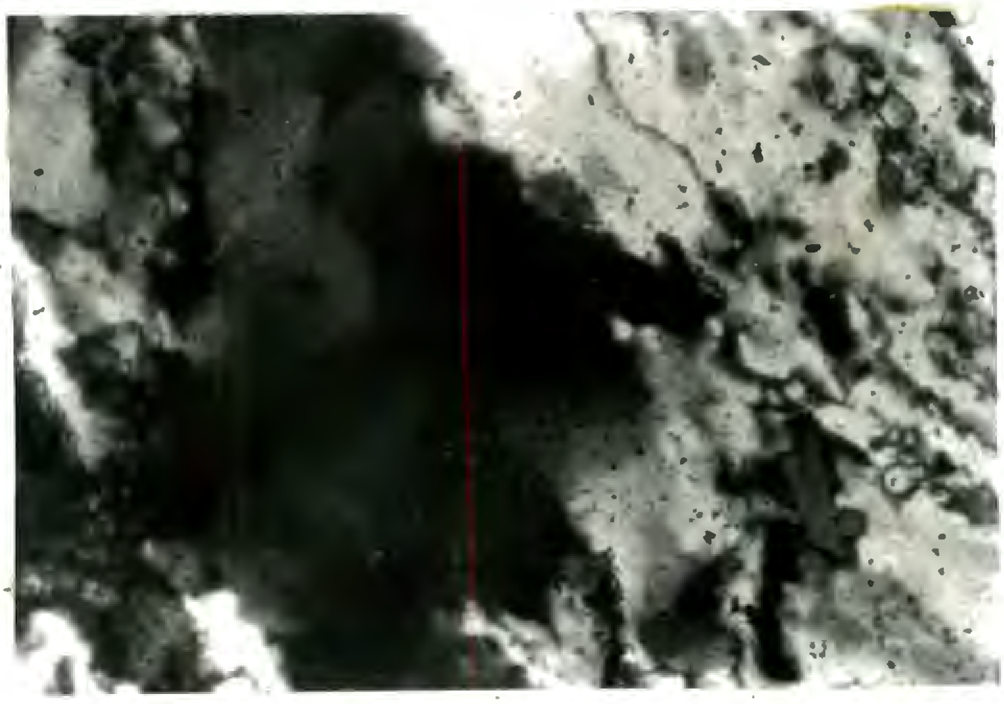


Plate 36. Photomicrograph showing sub-grain development in a quartz porphyroclast and it's envelope of mylonite grains. X100.

be composed of a mosaic of individual grains. The margins of these porphyroclasts are highly sutured and they are normally surrounded by a broad envelope of fine-grained quartz (Plate 36). Plagioclase is saussuritized or, more commonly, contains deformation twins. Flame varieties of perthite are spectacularly developed in microcline with the exsolution lamellae extending completely across the crystals (Plate 37). Biotite appears as deformed and shredded crystals and is normally reduced to chlorite. Hornblende shows surprisingly few effects (see also Johnson 1961) but is sometimes rimmed by chlorite and epidote.

Mylonites representing the final stage are quite obvious in the field. They are normally black or red, fine-grained rocks - even where developed in leucocratic gneisses. Porphyroclasts are still fairly common and these are associated with streaks and thin seams of lighter coloured material. Ultimately mylonitisation produces a finely comminuted rock flour which gives rise to a rock with alternating light and dark bands and overall flaggy appearance.

Under the microscope it can be seen that the original quartz grains have completely disappeared and the remaining porphyroclasts are usually ovoid plagioclase crystals with deformation twins or perthite.

3. *Structures of the mylonite belt*

Viewed as a regional feature the mylonite belt has a northwest trend (approximately 115^{o}) but within the study area one segment of the mylonite belt on Pelladrift Farm strikes east-west. This is revealed in Fig. 53 in which data from the mylonites are represented on three orientation diagrams representing three equal divisions of the belt in the mapped area.

The mylonite planar fabric dips uniformly north between 50^{o} and 80^{o}. There is more than one mylonite planar fabric represented in this belt and some zones can clearly be seen to crosscut others (Plate 38). One rare type of planar fabric is produced by the flattening of disoriented and angular hornblende-bearing rock fragments in the hornblende agmatites just north of the Lineament (see map). Where they become involved in the mylonite belt the fragments become drawn out and are composed of a retrograde assemblage of chlorite and epidote (Plate 39). Adjacent to the mylonite belt the linear fabric in the gneisses is approximately horizontal and trends parallel to the ZAHNCAFS. One of the first indications in the field that the Lineament is being approached is the noticeable undulation of this linear fabric on the foliation surface (Plate 21). Ramsay (1960) explained this phenomenon by assuming that the foliation containing an undulating lineation is contained within the 'ab' plane of simple shear (Fig. 54).

These lineations, i.e. pre-ZAHNCAFS fabrics, are found throughout the belt but a further lineation is developed on the mylonite foliation. This is a rectilinear feature and shows no evidence of undulation as do the earlier fabrics. Owing to the fine-grained nature of the rocks it is not always possible to precisely define this fabric and often it is only recognised by

Plate 37. Photomicrograph of a microcline grain showing the development of flame perthite. X40.

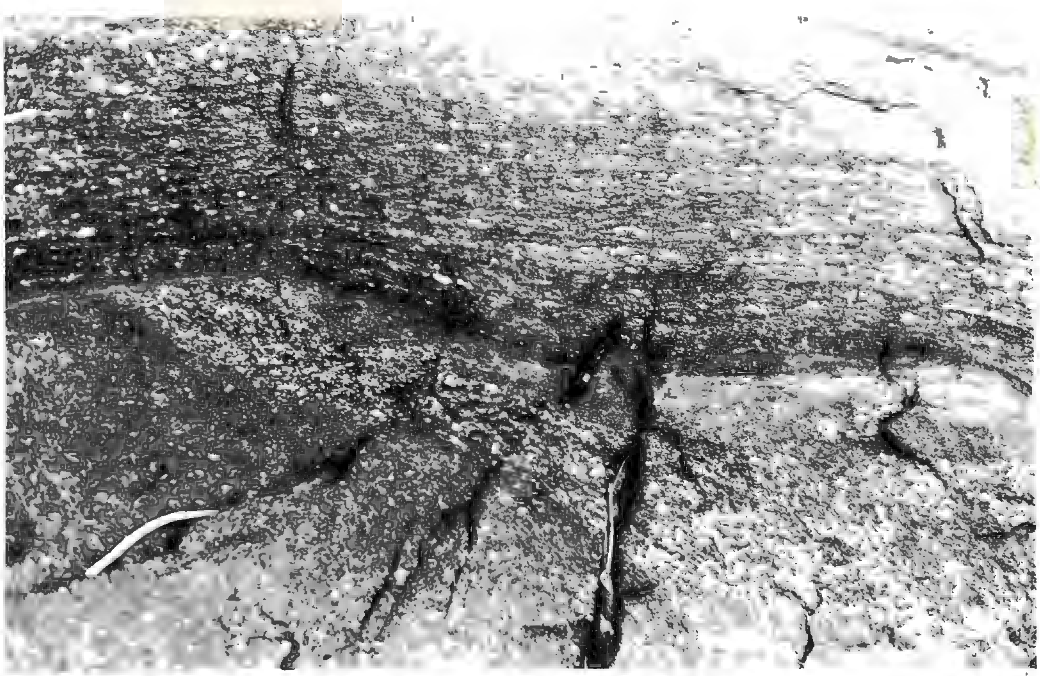


Plate 38. Two generations of mylonite in the Pofadder Lineament. Pelladrift Farm.

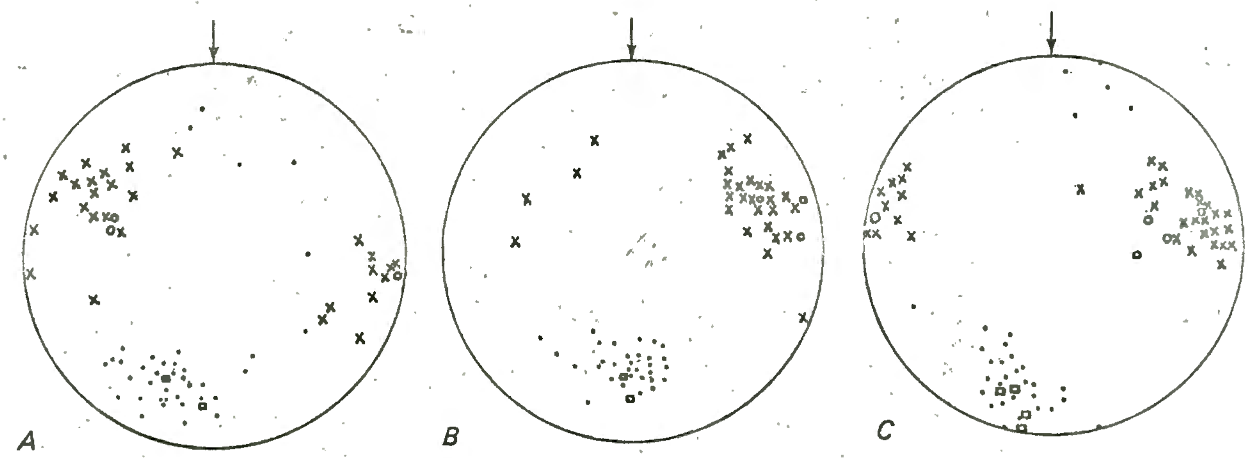


Figure 53. Orientation diagrams for data collected in the Pofadder Lineament. Each diagram represents approximately 7 km. A. Western section, 38 poles to foliation surfaces, 29 lineations, 3 fold axes, 2 poles to fold axial planes. B. Central section, 27 poles to foliation surfaces, 27 lineations, 2 fold axes, 2 poles to fold axial planes. C. Eastern section, 32 poles to foliation surfaces, 33 lineations, 4 fold axes, 4 poles to fold axial planes.

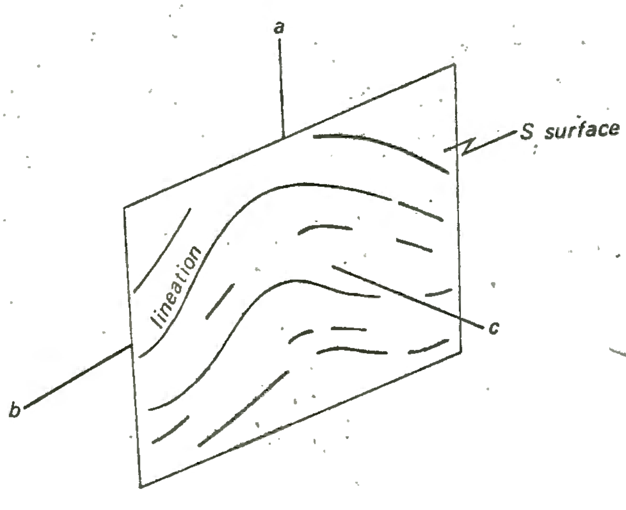


Figure 54. Variation of lineation plunge on a foliation surface where the surface is parallel to the 'ab' plane of simple shear.

a colour streaking on the foliation surfaces. Sometimes it is identifiable as streaks of quartz or quartz and feldspar. This lineation is indistinguishable in orientation from the earlier fabric and both are sub-horizontal or have a moderate plunge. Sometimes the lineation is identifiable as a surface intersection where the mylonite foliation cuts a previous planar fabric.

Folds can be found which are defined by the contacts between lithological layers and they appear to have the mylonite foliation as an axial plane cleavage (Plate 40). It was not possible to get suitably oriented photographs of these folds (c.f. chapter VI) and no quantitative classification could therefore be made, but many of them, appear to be similar folds. Everywhere such folds are present any linear fabric in the mylonite was found to be parallel to the fold axis and they are therefore B-lineations. A peculiarity of these folds is that they appear to mirror the orientation of pre-ZAHNCAFS mesoscopic folds in the surrounding gneiss. Occasionally both ZAHNCAFS folds and pre-ZAHNCAFS folds with identical orientations can be found in the same outcrop (Plate 41). No folds have been positively identified which deform the mylonite foliation. A considerable amount of controversy has arisen over the dynamic significance of mylonite structures (c.f. Johnson and Christie in discussion 1965) and it would appear that these structures have no unique interpretation. The kinematics of ZAHNCAFS are best explained by recourse to the Ramsay and Graham (1970) model and will be discussed in the next section. In one part of the mylonite belt there occurs an extensive development of pegmatite (Skimmelberg pegmatite - see map). Much of this pegmatite is cut by the mylonite foliation but in places the rock is entirely unstrained. No evidence of pegmatites cutting the mylonite fabric was found but undeformed areas suggest that, at least in part, they may postdate the mylonite. It is apparent both in the field and from air photograph interpretation that these pegmatites trend north-south and form an en-echelon pattern in the mylonites. These arrays of tension fissures are the result of simple shear and they form parallel to the direction of maximum shortening (the Z-direction) in the shear zone (Fig. 55). The direction of movement in the ZAHNCAFS is responsible for their orientation and in the Pofadder Lineament they confirm the dextral displacement.

Small scale examples of tension fissures have been encountered in mylonite fabrics (Plate 42) and they provide supporting evidence for the complex nature of the mylonite development. Small fractures, sometimes with slickensides, crossing the mylonites are fairly common. They can also be found throughout the Onseepkans area and there is no noticeable increase in their frequency in the Pofadder Lineament. For this reason they are not thought to be genetically related to the formation of the ZAHNCAFS although such fractures would be expected in mylonites because, being weak rocks, they localise subsequent strains (c.f. section VB1).

It has been suggested by Beukes (1973) and Joubert (1974, p.20) that the mafic rocks found within the confines of the mylonite belt were intruded during or after the formation of the ZAHNCAFS. These conclusions are difficult to assess since neither author cites the evidence on which the interpretations were made. The main mafic body to fall in this category in the Warmbad area is the Tantalite Valley Metagabbro (Beukes op. cit., p.243-253) but where

Plate 39. The flattening of angular hornblende-rich fragments to form a planar fabric. Compare with Plate 2. Pelladrift Farm.

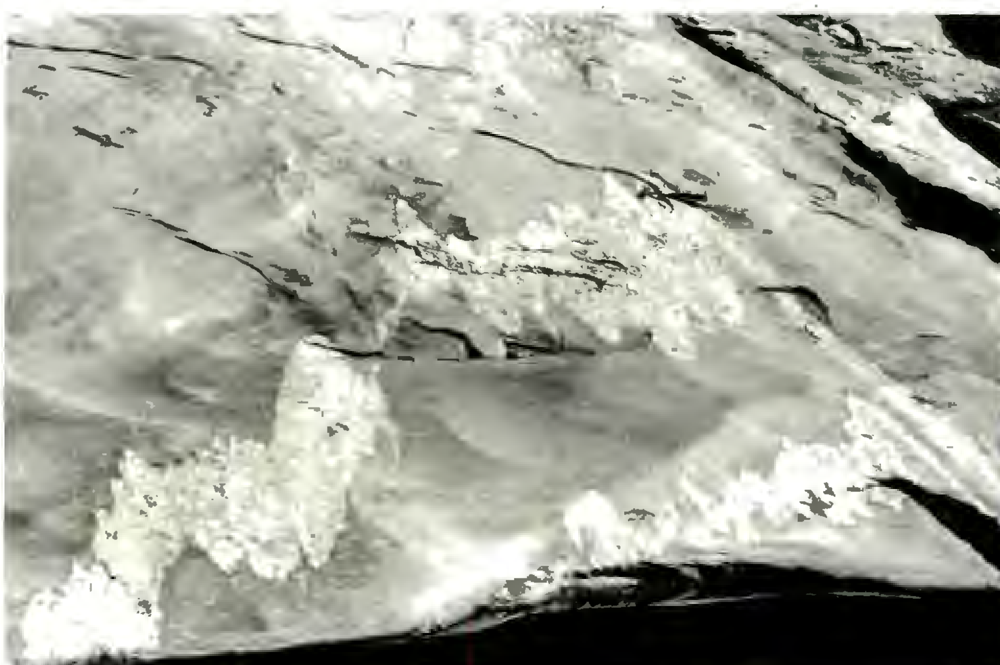


Plate 40. Folds defined by pegmatites in the Pofadder Lineament with the mylonite foliation as an axial plane cleavage. Kum Kum Farm.

Plate 41. Folds in mylonites (lower left) with an identical orientation to folds in the surrounding gneisses (right hand side) Kum Kum Farm.

Plate 42. Zone of tension fissures developed in mylonite. Kum Kum Farm.

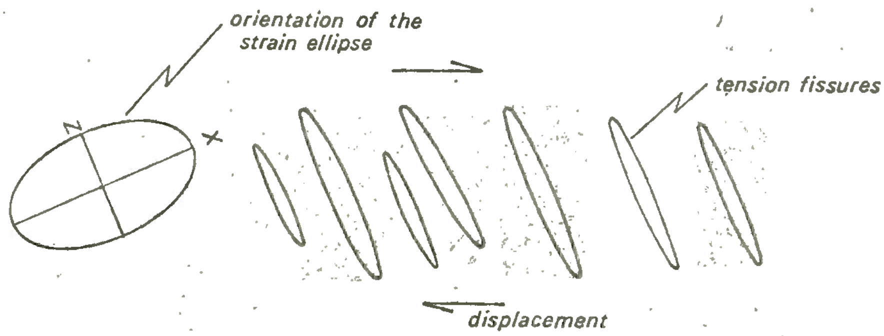
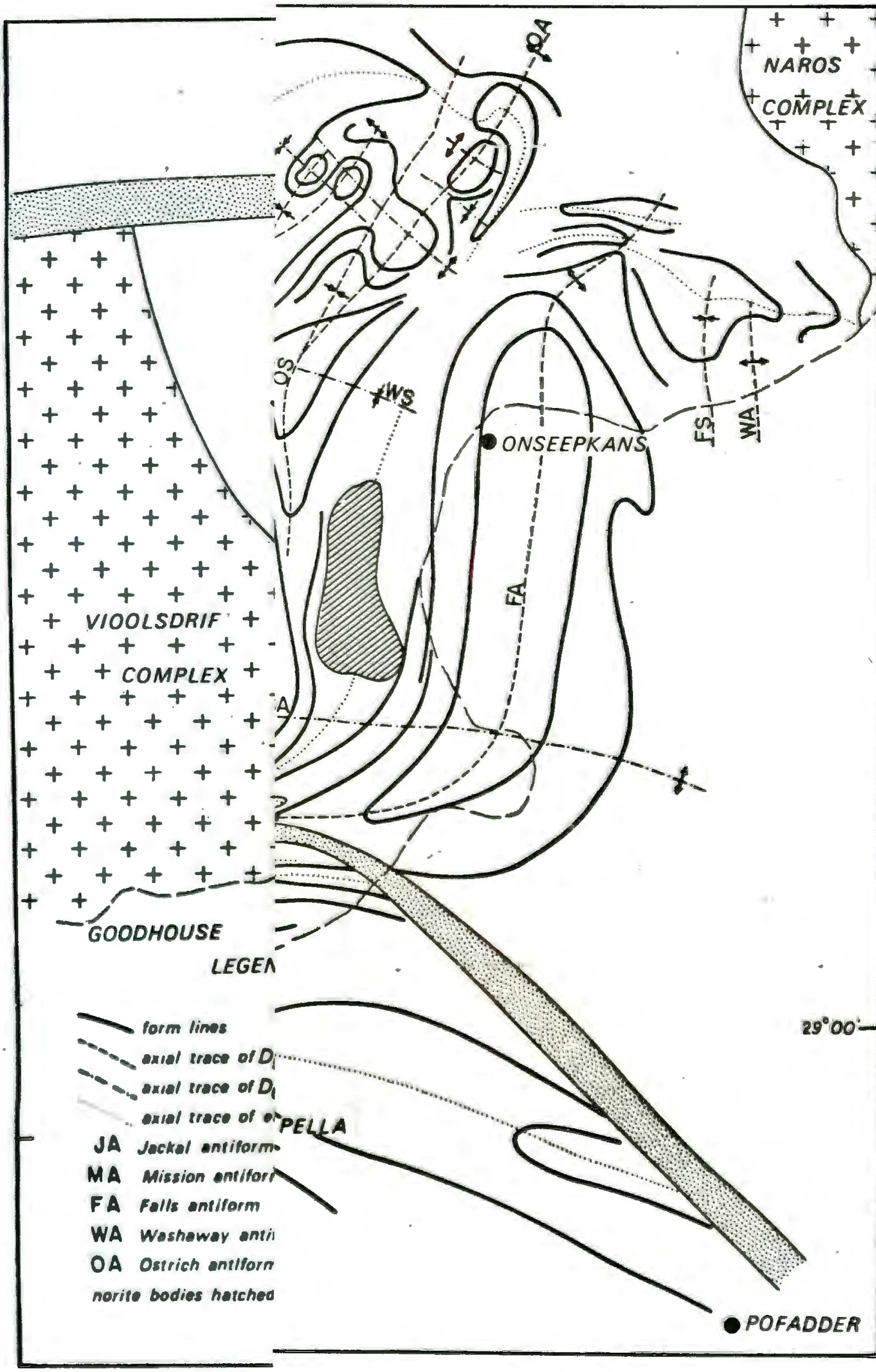


Figure 55. Diagram showing the relationship of tension fissures to the simple shear strain ellipsoid and the displacement in shear zones.

examined by the writer the marginal zone of this body is strongly refoliated and retrogressed. No intrusive contact between these mafic bodies and the mylonites have been recorded and neither of the above authors reports mylonite xenoliths in the mafic rocks. The absence of mylonitisation in the cores of these bodies has, perhaps, been the main criterion for their postulating age. It was shown in Chapter IV, however, that the behaviour of isotropic rigid bodies during deformation was quite different from that of anisotropic layered rocks and the penetrative foliation, by whatever mechanism, in these bodies is not normal. The Tantalite Valley body is an orthopyroxene-bearing mafic rock that appears to be indistinguishable in composition (Beukes, 1973, p.248) from similar rocks found throughout the Warmbad-Onseepkans area (Moore 1976) which are described in this report as norites of the charnockitic rock suite.

4. *Zone of reorientation*

The re-orientation of the gneisses in the northern block can be seen on the accompanying geological map and can readily be appreciated on the ERTS image of the area (Plate 43) where dextral displacement is clearly defined. It can be seen on this photograph that the reorientation pattern is visibly different in the northern and southern blocks. To the south of the Lineament there is a wide zone where the orientation of the foliation in the gneisses is parallel to the Lineament whereas in the northern block there is a more abrupt change in trend away from the mylonite and this is consistent over the entire area covered by the photograph. The structural pattern in the northern block is connected with the development of a series of large crescent-shaped antiforms or elongate domes (D_5), from west to east, the Jackal, Mission and Falls antiforms (Fig. 56). The Falls antiform in particular is very large and its axial trace can be followed for 40 km.



NAROS
COMPLEX
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FA
VIOOLSDRIF
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FA Falls antiform
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29°00'
POFADDER

These folds are also present north of the Lineament in the Warmbad area. Beukes (1973) interpreted this difference in style as being due to displacement which brought into contact areas of dissimilar structure. However, the writer believes there are unique features about these folds that strongly suggest a genetic connection with the mylonite belt.

All these antiforms or dome-shaped structures are doubly plunging. Their deformation defines an antiformal axial trace (Shining Sister antiform) trending parallel to the Lineament and producing an H-type fold interference pattern (Fig. 25), referred to here as D_6. These folds have a unique spatial relationship with the mylonite. Adjacent to the Lineament the fold axial planes of the crescent-shaped antiforms trend northwest and the folds plunge gently northwest; in this area the folds are tight-to isoclinal. Away from the mylonites the fold axial planes change rapidly in trend from northwest to northeast and the folds open out and achieve their greatest wavelength. The axial trace of the Shining Sister antiform crosses the folds at this point and thereafter they achieve, more gradually than before, a northerly trend and become tighter. At their northern extremities they close and plunge approximately north but in this area their dihedral angle is greater than that seen adjacent to the Lineament. Two crescent-shaped antiforms north of the Lineament at Tantalite Valley have an identical style (Beukes 1973, geological map). It should be noted that all these folds are similar in size and are not truncated by the mylonites but close completely just north of the Lineament. The structural pattern both north and south of the Pofadder Lineament is illustrated in Fig. 56. The intervening synforms in type-H fold interference patterns are generally obscure and only an inferred axial trace can be placed between the Jackal and Mission antiforms. The O'Connell synform separating the Mission and Falls antiforms is quite evident and can be traced for some distance to the north where it dies out in a series of open dome and basin structures. The northwesterly trending fold direction whose presence produces the dome and basin structures appears to be related to the similar-trending Shining Sister antiform. The D_5 and D_6 folds in the Pofadder ZAHN-CAFS have produced one of the best preserved examples of a large scale type-H interference patterns recorded in the literature. It has been shown in chapter IV that these folds deform two previous generations of structures (D_3 and D_4) in the western sub-area and there is no evidence that D_5 and D_6 folds extend outside the zone adjacent to the Lineament in the northern block.

South of the Lineament these folds cannot be recognised. There are no unique interference patterns typical of this area and earlier folds are merely brought in parallelism as they approach the Lineament and are then truncated by the mylonites (Fig. 56). This is not only the case in the Onseepkans area but is also true for the folds shown by Beukes (1973) in the Warmbad area and at Pofadder a major synform with a quartzite core shown on maps by Joubert (in preparation) is also truncated by the mylonites. None of these folds is doubly plunging and no dome and basin structures or crescent-shaped folds of any type exist.

It is interesting to note that in the Ai-Ais area some 100 km farther northwest, a similar disparity in structural style has been noted across the Pofadder Lineament. A zone south of the Lineament in which structural trends

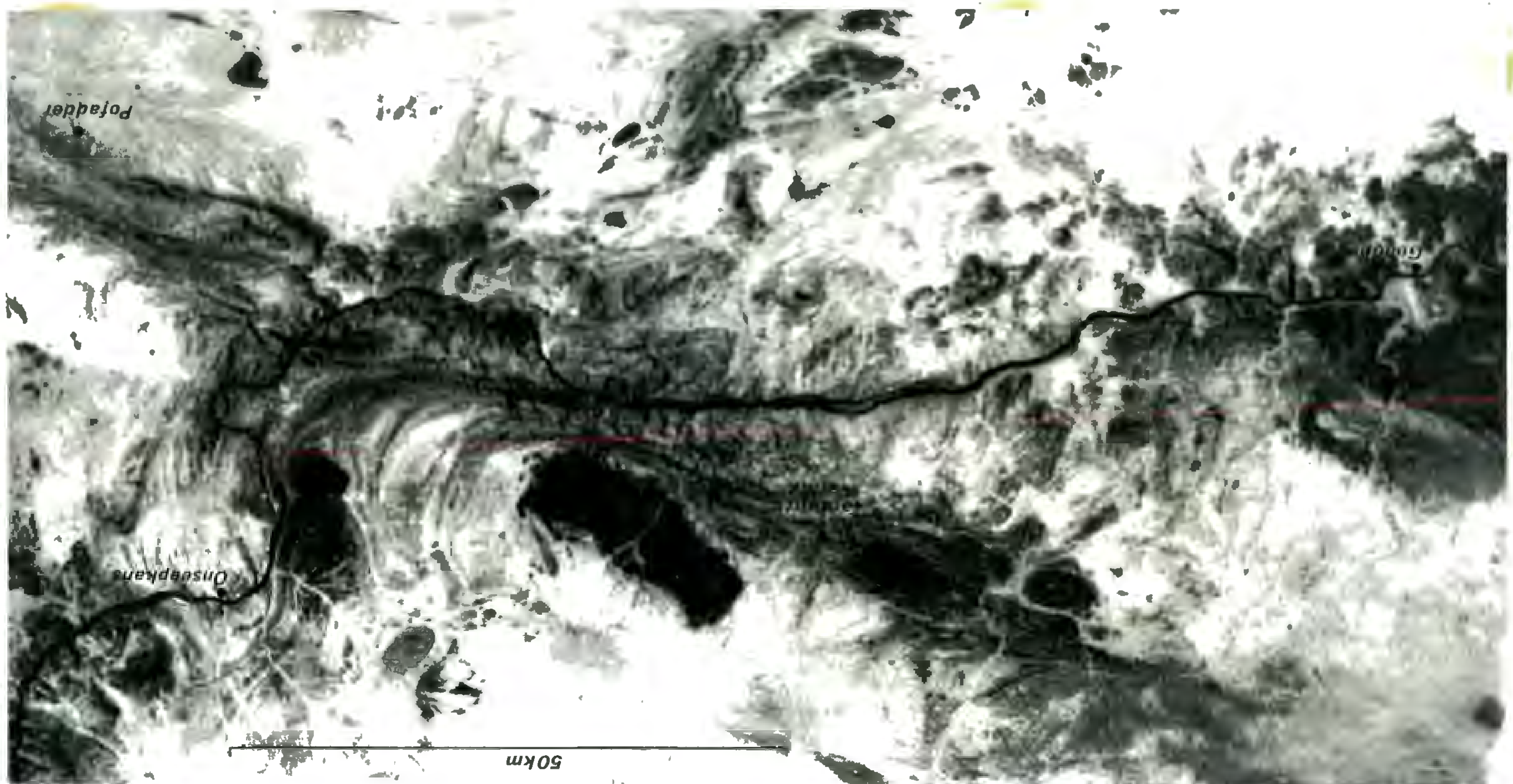


Plate 43. Part of an ERTS photograph showing the area between Goodhouse and Onseepkans. Note the Pofadder Lineament extending from upper left to lower right through Tantalite Valley.

are parallel to the shear direction has been described by Blignault et al. (1974) and large macroscopic folds are reported adjacent to the Lineament in the northern block. Unfortunately the zone of parallel alignment in the southern block is sandwiched between the Vioolsdrif complex and the mylonites which has led Blignault et al. (1974, p. 45) to misinterpret it as a 'marginal zone' of the Namaqua belt, unconnected with the development of the shear zone.

It has therefore been independently confirmed by three separate workers that the structural pattern differs radically across the Pofadder Lineament.

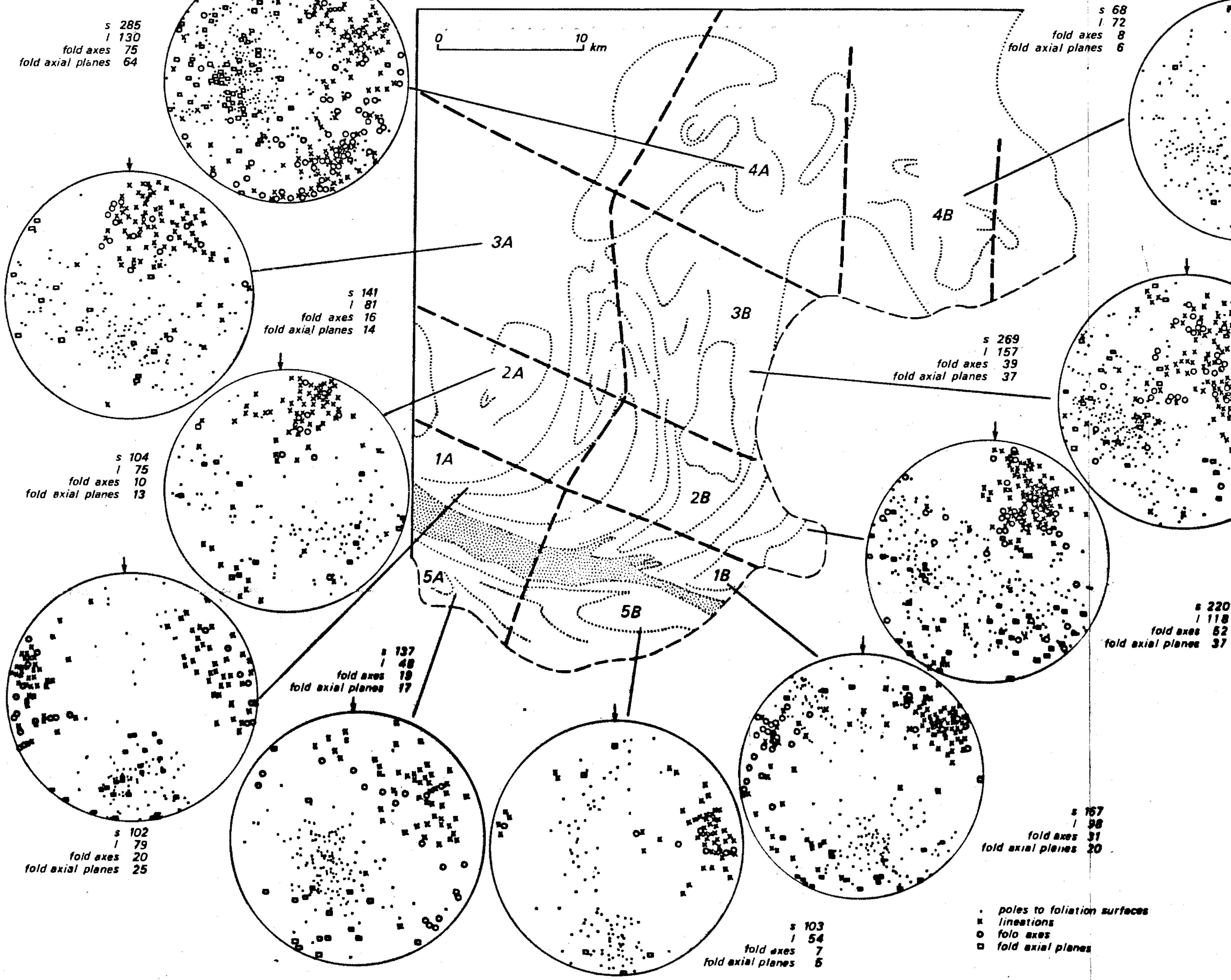
5. *Fabric elements associated with D_5 and D_6 folds and the reorientation of earlier fabric elements and folds*

No penetrative fabric elements have been found associated with either D_5 or D_6 structures. Mesoscopic folds are encountered in the hinge zones of the Mission and Falls antiforms adjacent to the Lineament and also in the hinge of the O'Connell synform (well exposed to the south of the road just west of the homestead on Keimasmond Farm). In the core of the Mission antiform it has already been noted that a mylonite foliation is locally present, axial planar to the fold. In the core of the Falls antiform adjacent to the Lineament narrow pegmatite bands are found parallel to the axial planes of minor parasitic folds. At the northern boundary of the ZAHNCAFS it can be demonstrated that the D_3/D_5 relationship is one of coaxial refolding and only a slight difference is noted in the l_3 azimuth around the hinges of the D_5 folds. Because of the modification of fold shapes adjacent to the Lineament, this slight variation is no longer discernable and it appears here that the easterly or westerly plunging l_3 lineations and D_5 folds are related structures.

At the northern hinge zone of the Mission antiform, for example, the plunge of l_3 is approximately north (zone 3A, Fig. 57) but traced southwards the plunge decreases in amount until, approximately at the axial trace of the Shining Sister antiform it is horizontal (see map). South of the axial trace the trend changes to a shallow easterly or westerly plunge as the mylonite belt is approached (zone 1A, Fig. 57). A similar change in l_3 orientation takes place along the Falls antiform but l_3 in the hinge zone of the D_3 Leopard synform, while changing in orientation, does not develop a westerly plunge as the Lineament is approached but merely plunges at a shallower angle to the east-northeast (zone 1B, Fig. 57). Since the shear strains are higher adjacent to the Lineament than at the boundary there has been a more marked rotation of the fabric elements towards the XY plane of the simple shear strain ellipsoid. When comparing the orientation diagrams of the zones adjacent to the Lineament (zones 1A and 1B, Fig. 57) with those at the margin (especially zones 4A and 3B, Fig. 57) it can be seen that the latter display a much lower fabric symmetry.

A locally complex fold pattern is produced by the $D_4/D_5/D_6$ interference at the boundary of the ZAHNCAFS. Domes and basins are formed by D_6 folds being superimposed on D_5 folds, notably domes at the intersection of the First Camp and Ostrich antiforms (c.f. Annexure 2). This superposition of the D_5 O'Connell and Ostrich structures on D_4 folds has produced a type-H interference

Fig.57 Orientation diagrams for data collected in the western sub-area.
The sub-area is divided into zones trending parallel to the Pofadder Lineament.



pattern (Fig. 25). The D_4 Fortress and Bushman Hills antiforms probably belong to a single structure which is now discontinuous because of the refolding effect of the O'Connell synform. The l_3 lineation pattern in this area is highly irregular.

B. KINEMATIC INTERPRETATION OF THE POFADDER ZAHNCAFS

1. *The mylonites*

It was suggested by Johnson (1967) and Ramsay (1967, p.89, Fig. 3-26) that mylonite foliation is not a surface of simple shear. This hypothesis was subsequently proved by Ramsay and Graham (1970) who confined their studies to isotropic rocks thereby avoiding the confusion introduced by pre-existing anisotropies. They showed that mylonite foliation is perpendicular to the direction of greatest shortening in the rock, that is, parallel to the XY-plane of the strain ellipsoid. As stated earlier they have also demonstrated that the S-shaped pattern across shear zones is not the result of drag but due to the expansion of the shear zone during progressive deformation by heterogeneous simple shear.

Work on mylonites in recent years has produced interpretations that diverge strongly from the classical view that they are the result of cataclasis, i.e. the mechanical degradation and crushing or brecciation of rocks at low temperatures. Much of this recent work appears to be based on the experiments of Carter et al. (1964) which showed that typical 'cataclastic' textures can be produced by plastic deformation and recrystallisation of quartz. Katz (1968) pointed out the implications of these findings and suggested that mylonites were the result of ductile deformation.

Since then several workers have stressed the ductile nature of mylonite formation (Bell and Etheridge 1973; Moore 1973; Vernon 1974) but undoubtedly the most significant contributions in this field have been those of White (1973, 1975b, 1975c) and White and Treagus (1975) who based their conclusions on similar work done in the fields of ceramics and metallurgy. The quartz textures seen in the Pofadder mylonites are very similar to textures described in these studies and lead to the conclusion that the Pofadder mylonites have been formed by a process of dislocation and recovery that are essential characteristics of hot working, i.e. a dynamic process occuring during deformation (White and Treagus op. cit., p.220).

White (1975b, p.246) has pointed out the essential similarity between the quartz fabric in normal metamorphic tectonites and mylonites and has suggested that the latter may indicate higher strain rate deformation. The elongated quartz grains with undulatory extinction seen at the onset of mylonitisation indicate high densities of tangled and matted dislocations and are the result of strain hardening processes in an initial stage of primary creep. As the dislocation density increases, the internal strain energy of quartz is increased and elongate subgrains form producing the deformation bands seen under the microscope. This is thought to be the start of a process of dynamic recovery and marks the transition of the fabric to a second stage of steady state or secondary creep. These elongate sub-grains tend to become equidimensional with high angle boundaries as deformation increases and this produces the mosaic effect in quartz seen in more advanced stages of mylonitization.

It is worth noting here that these strain-free sub-grains may be sufficiently misoriented from their neighbours to appear as individual entities, thus giving rise to a 'mortar texture'. White (1973) has noted that the sub-grains can be confused with new strain-free nuclei and has pointed out that the distinction can only be made with the aid of electron microscopy. The writer considers this highly significant in view of the classic interpretation of this texture as defining blastomylonites. In the last stages of the mylonitisation process the rock is completely reduced to a finer-grained product than the original rock with the recrystallised grains derived from the misoriented sub-grains. This process is illustrated diagrammatically in Fig. 58.

White (1975b) has defined two types of mylonite and distinguished a third rock type which he refers to as a cataclasite. In the first type of mylonite the porphyroclasts and finer mylonite grains show no evidence of recovery and recrystallisation alone has taken place. In the second category sub-grains are abundant and the mylonite forms by dynamic recovery. These two types are thought to form at moderate strain rates (around 10^{-5} to 10^{-6}/sec). The cataclasites represent the classic case where porphyroclasts form by mechanical degradation. There are very few sub-grains and there is no evidence of recovery. The strain rates produced in cataclasites are thought to be greater than 10^{-4}/sec. In this classification the Pofadder mylonites would fall in the second category.

The formation of the Pofadder mylonite belt by a process of hot working or creep is also more compatible with other evidence. The reorientation of the gneisses surrounding the Pofadder Lineament is adequately seen on the geological map and is spectacularly demonstrated on ERTS photographs of the region (Plate 43). These reoriented high-grade and medium-grade gneisses show no signs of retrogression and it would appear self evident that the deformation took place at elevated P-T conditions.

Small ZAHNCAFS in these gneisses are ubiquitous and become more abundant as the Lineament is approached. They vary from a few centimetres to several metres in length and are identical to similar zones in the Lineament both in size, orientation and displacement. A notable feature of some ZAHNCAFS is that they appear to have formed under conditions of anatexis and small biotite-bearing leucosomes are found in their cores (Plate 44). By far the greater majority of the ZAHNCAFS have no refoliated rocks, i.e. mylonites, in their cores and they simply occur as small monoclinal folds (Plate 45). No evidence of retrograde metamorphism has been found in these structures. This appears to be strong supporting evidence for the assumption of high P-T conditions during the formation of the Pofadder ZAHNCAFS. It was found that the minor ZAHNCAFS are much more prevalent north of the Lineament than to the south. This may be due to the limited extent of the southern block seen by the writer

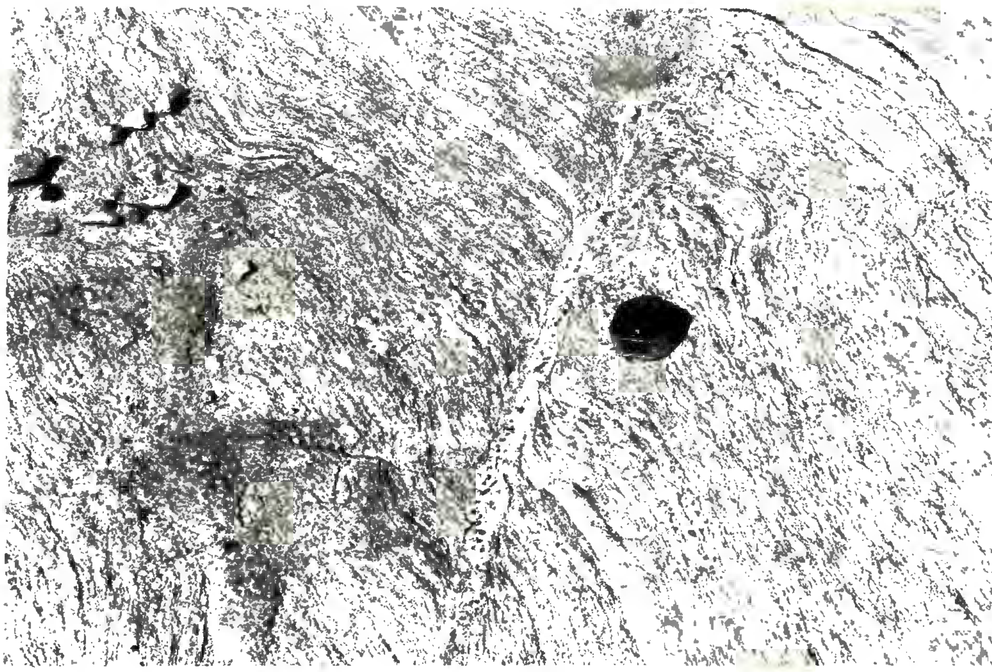


Plate 44. Minor ZAHNCAFS showing leucosome formation in the core. Keimasmond Farm.

Plate 45. Minor ZAHNCAFS represented by folds with little or no accompanying refoliation. Keimasmond Farm.

in the mapped area but it may also be due to higher P-T conditions indicated by the metamorphic grade of the rocks in the northern block (c.f. chapter III).

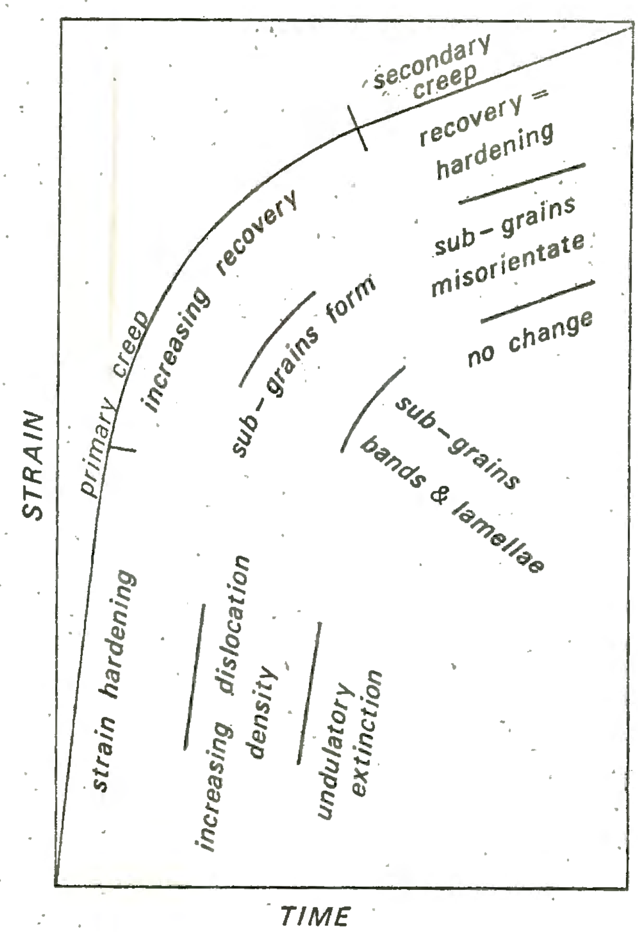


Figure 58. The deformation mechanisms and optical features of quartz produced during creep. After White and Treagus (1975).

Shear zones and mylonite belts are normally associated with non-penetrative deformation and in most texts are almost exclusively discussed as single, linear features. However, this is not always the case. The 'Laxford Front' in the Scottish Highlands is characterised by a suite of such shear zones and it would appear that Laxfordian metamorphism and deformation is the result of this style of deformation becoming penetrative over wider areas (Sutton and Watson 1962; Beach et al. 1974). Katz (1968) also describes mylonitic textures that are penetrative over wider areas and are not confined to a single zone. The Pofadder Lineament is an example of the more common case where the effects are limited in lateral extent. The mineralogy of the rocks in the Lineament indicate that low-grade metamorphic conditions were dominant (Winkler 1974, pp.73-81) although one sample (D.T. 1093, Plate 46) at the margin of the main zone showed apparently stable sillimanite. This example is unique and generally the presence of chlorite (formed at the expense of biotite and hornblende) and muscovite and the absence of cordierite, sillimanite and garnet define the prevailing grade.

If high P-T conditions are proposed for the formation of the Pofadder ZAHNCAFS why is it then that mylonite zones in the core of the structure (and in many others) indicate low-grade conditions? The answer to this problem has been partially given by Watterson (1975) who pointed out that many lineaments remain active over very long periods of time - up to 1000 Ma in some cases (op. cit., p.521), and he suggested that these are zones of inherently weak rocks that localise any strain affecting a wider area. This interpretation is supported by the following relationships:

$$d \propto \left(\frac{\sigma}{\mu}\right)^{-m} \qquad \text{(White 1975b, p.580)}$$

Where μ = shear modulus
and m = constant of value 0,5

and $\dot{e} \propto \sigma^2$ (White 1975b, p.581)

and $\dot{e} \propto d^{-1}$ (Hahn et al., 1967, p.776)

It is obvious that an accelleration of strain rates is to be expected in the core of these ZAHNCAFS since these structures grow in width during progressive deformation and therefore reduction in grain size is most pronounced in the core. White (1975c) states that sub-grains in mylonites are between 2-10 μm in diameter whereas in normal tectonites they may vary between 10-25 μm in diameter and has suggested that

$$\left(\frac{d \text{ mylonite}}{d \text{ metamorphite}}\right)^2 = \frac{\sigma \text{ metamorphite}}{\sigma \text{ mylonite}} \approx 10 \qquad \text{(White 1975, p.580)}$$

The writer therefore suggests that lineaments are active through long periods of geological time and are also active at comparatively low crustal levels where P-T conditions favour low-grade metamorphic assemblages. It has been shown by Beach and Fyfe (1972) that lineaments act as channel ways for migrating water and metamorphic reactions will consequently be facilitated. This interpretation may be generally applicable to other mylonite belts but examples of mylonites containing high-grade assemblages are not uncommon (Sutton and Watson 1959, 1962; Theodore 1972; Lundgren and Ebblin 1972; Bak et al. 1975a and 1975b; Beach 1974).

The metamorphic grade now indicated in the Pofadder Lineament does not, therefore, reflect the initial grade of deformation in the ZAHNCAFS but is a comparatively late feature.

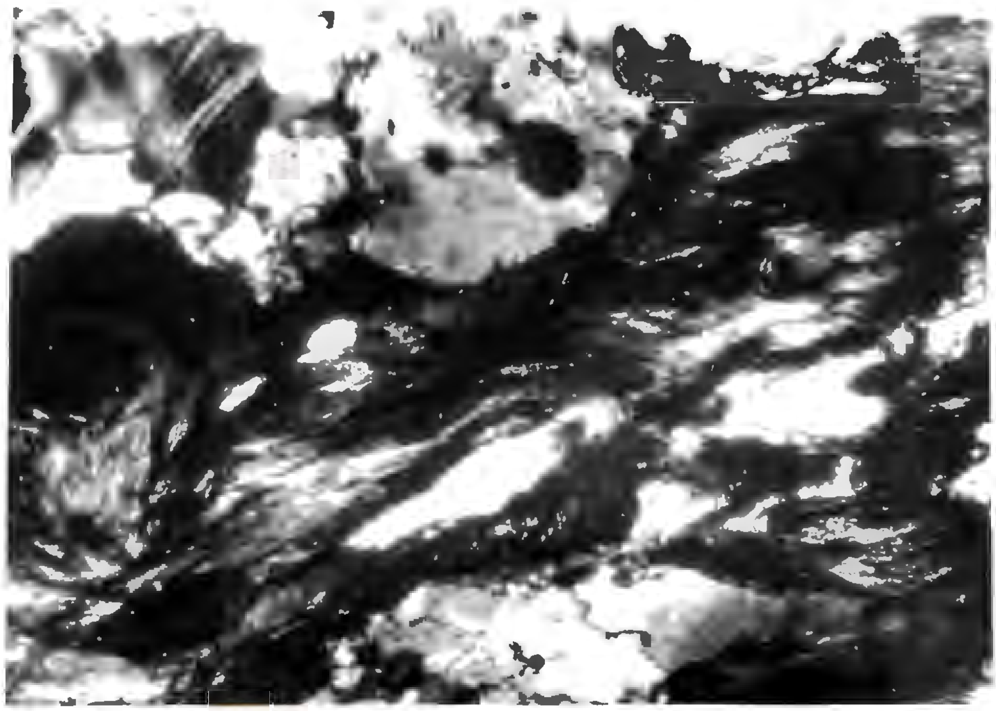


Plate 46. Photomicrograph showing apparently stable sillimanite in a mylonite fabric. X40.

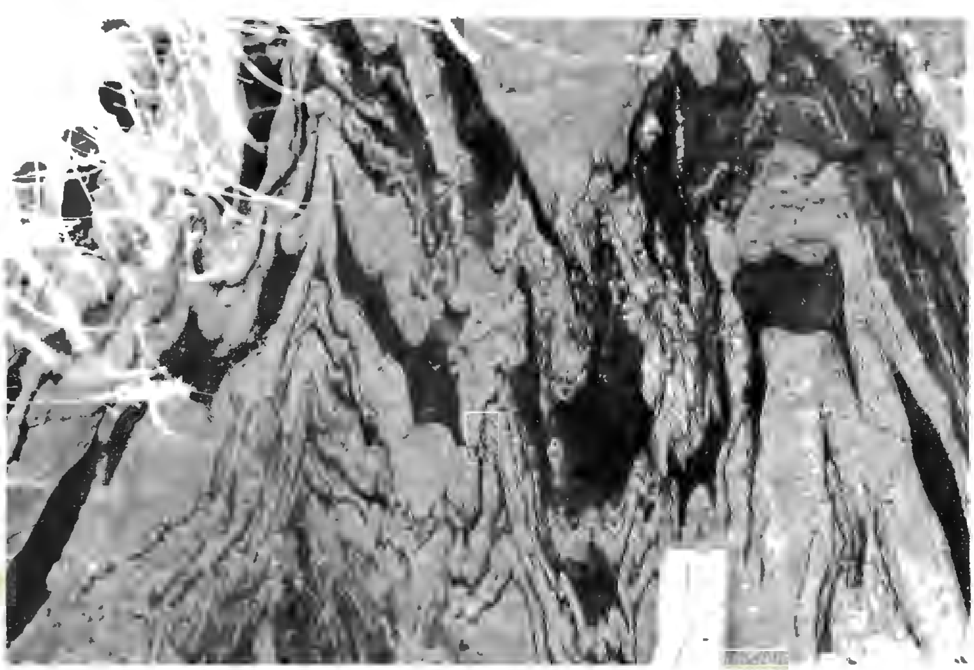


Plate 47. Folds defined by migmatite neosomes where the anastomosing nature of the neosomes is obvious. Keimasmond Farm.

The recognition of high P-T conditions during the formation of the ZAHNCAFS has important implications in the light of a recent publication by Joubert (1974a) in which the Pofadder Lineament is interpreted as a wrench fault in terms of the Moody and Hill (1956) wrench fault tectonic model. This model is based on the Anderson (1951) fault classification system where faults are analysed dynamically by Navier-Coulomb fracture criteria - generally accepted as being valid only for the brittle deformation of rocks (Ode 1960; Price 1966; Jaeger and Cook 1969). Since it has been shown that the Pofadder Lineament results from ductile deformation an interpretation involving assumptions of brittle failure would seem to be questionable.

It may also be asked whether the brittle behaviour of rocks is still a tenable hypothesis at these postulated P-T conditions (650°C between 3,5 and 5 kb in the southern block and 650° to 700°C between 5 and 7 kb in the northern block: see Fig. 23). According to Griggs and Handin (1960, p.359) all rocks investigated experimentally at 500°C and 5 kb showed ductile behaviour with only one exception. According to Orowan (1960) brittle deformation is limited to the upper 10 km of the crust (∿ 3 kb confining pressure). Edmund and Murrell (1973) placed a maximum confining pressure of 7 kb at room temperature on the brittle behaviour of rock. This 7 kb figure will be considerably reduced if higher temperatures are taken into account as the yield point between brittle and ductile deformation will be lowered (Heard 1960).

2. *Zone of reorientation*

Since the publication of the Ramsay and Graham (1970) paper the simple shear model has been successfully applied in many areas at all scales of observation. Shear zones on the scale of the Pofadder ZAHNCAFS have been described in Greenland by Bak et al. (1975a, 1975b) and in Canada by Garnett and Brown (1973). The model has also been used to explain the origin of the Laxford Front in the Scottish Highlands (Beach 1974) and it may also be extended to the scale of mobile belts. Bridgwater et al. (1973) have evoked the model to explain the deformation pattern in the Nagssugtoqidian mobile belt in Greenland and Coward et al. (1973) used to explain the deformation pattern in the northern margin of the Limpopo Mobile Belt in southern Africa. Escher and Watterson (1974) have proposed that the deformational features of the Grenville Front in Canada described by Dalziel et al. (1969) may also be interpreted as a zone of simple shear. The 'zones of straightening' described in the Mozambique Belt by Hepworth (1967, 1972) have been interpreted by Escher and Watterson (op cit., pp. 226-227) in a similar manner. At the other end of the scale zones of anomolously-high strain several hundred metres in width have been interpreted in terms of the Ramsay/Graham model by Coward (1973a, p.126). Such narrow high-strain zones may be analagous to the 'shear zones' represented by the limbs of flexural flow folds described in Chapter IV (c.f. Fig. 43).

The recognition of the D_5 and D_6 folds in the northern block, their unique style, orientation, and spatial relationship to the Lineament raises the problem of finding a mechanism that can account for them. In the model proposed by Ramsay and Graham (1970) for the formation of shear zones, no folds were developed but this is almost certainly an atypical case since the rocks in question were isotropic.

Simple shear model experiments with layered media involving the production of folds have been made by Ghosh (1966) and Wilcox et al. (1973) although those made by the latter workers are rather elementary. In the writer's opinion a combination of the above experimental approach with the Ramsay/Graham heterogenous simple shear model can account for most of the features seen in the Pofadder ZAHNCAFS.

Ghosh (1966) and Wilcox et al. (1973) have shown that simple shear model tests with layered media almost always result in the production of folds but Ghosh (op. cit.) has shown that their orientation and development is entirely dependent upon the original relationship of the layers to the simple shear strain ellipsoid. In the Pofadder ZAHNCAFS the axis of simple shear (the shear direction) as deduced from the displacement pattern trends parallel to the Lineament and is approximately horizontal. The b axis (the line at right angles to a in the plane of shear) plunges steeply north and the c axis (the line at right angles to the ab plane) trends northeast and is approximately horizontal.

The D_5 folds in the ZAHNCAFS have their axial planes parallel to the b axis and their axes are contained approximately in the ac plane. In simple shear model tests folds with this orientation form when the layers are originally stacked parallel to the ac plane. At the initiation of deformation a series of en-echelon folds will be formed whose axial planes are parallel to the b axis and make an angle of 45° with the a axis (Fig. 59).

This 45° trend is also the orientation of the maximum extension direction (X-direction) in the rock at the first increment of shear strain (Ramsay 1967, Fig. 3-22; Ramsay and Graham 1970, Fig. 8) and the axial plane represents the XY-plane of the simple shear strain ellipsoid, i.e. it is forming perpendicular to the maximum shortening direction in the rock (Fig. 59). The orientation of the axial planes of these folds therefore corresponds to the foliation direction described by Ramsay and Graham (1970) in isotropic rocks.

At this stage it is thought advisable to discuss the further development of the folds in terms of the Ramsay/Graham theory since some of the experimental model tests involve static boundary conditions and an essential feature of the Ramsay/Graham theory is that the affected zone increases in width during progressive deformation.

As the shear zone widens during deformation the folds continue to grow away from the core in a direction parallel to their trend. At the margins of this expanding zone the strain parameters are those seen in the core of the zone at the initiation of deformation which result in fold axial planes oriented at 45° to the shear direction. However, in the core of the zone the first-formed segments of the fold will be rotated so that their axial planes take up an orientation closer to the a direction of simple shear. This is because all lines and planes within a zone of simple shear move towards the shear direction with increasing shear strain and, as a result of the boundary conditions of the model, the highest shear strains are found in the core of the ZAHNCAFS the relationship between shear strain and line orientation is illustrated in

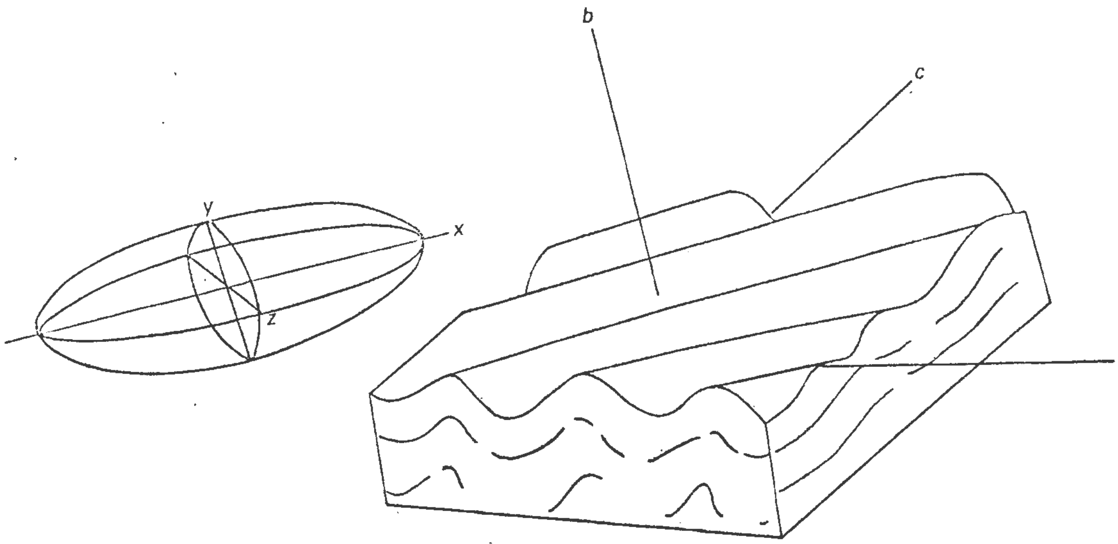


Figure 59. Orientation of folds formed at the first increment of strain during simple shear deformation where the layers are parallel to the ac plane. Inset shows orientation of the simple shear strain ellipsoid.

Fig. 60.

Thus far it has been shown that a series of en-echelon folds may be formed in the shear zone and that their axial surfaces will become curved as the shear zone grows in width. One further phenomenon of simple shear deformation is that the folds will not only become rotated but will become tightened in the core of the ZAHNCAFS. This can be seen by examining the relationship between shear strain and the ratio of the principal axes of the finite strain ellipse (Ramsay 1961, Figs. 3-20 and 3-21) as shown here in Fig. 61. The exact relationship is given by the equation:

$$\lambda_1 \text{ or } \lambda_2 = r^2 = \frac{\gamma^2 + 2 \pm \gamma(\gamma^2 + 4)^{\frac{1}{2}}}{2} \quad \text{(Ramsay 1967, p.84)}$$

Where r is the radius of the circle touching the ellipse

It is obvious that even for comparatively low values of shear strain ($\gamma = 5$) the ratio of $\lambda_1 : \lambda_2$ is approximately 25:1 and the rocks will have suffered some 80% shortening. The ratio of the principal strain axes escalates rapidly with increasing shear strain and, therefore, a marked change will be visible across the shear zone both in the total shortening suffered by the rocks and the dihedral angle between fold limbs.

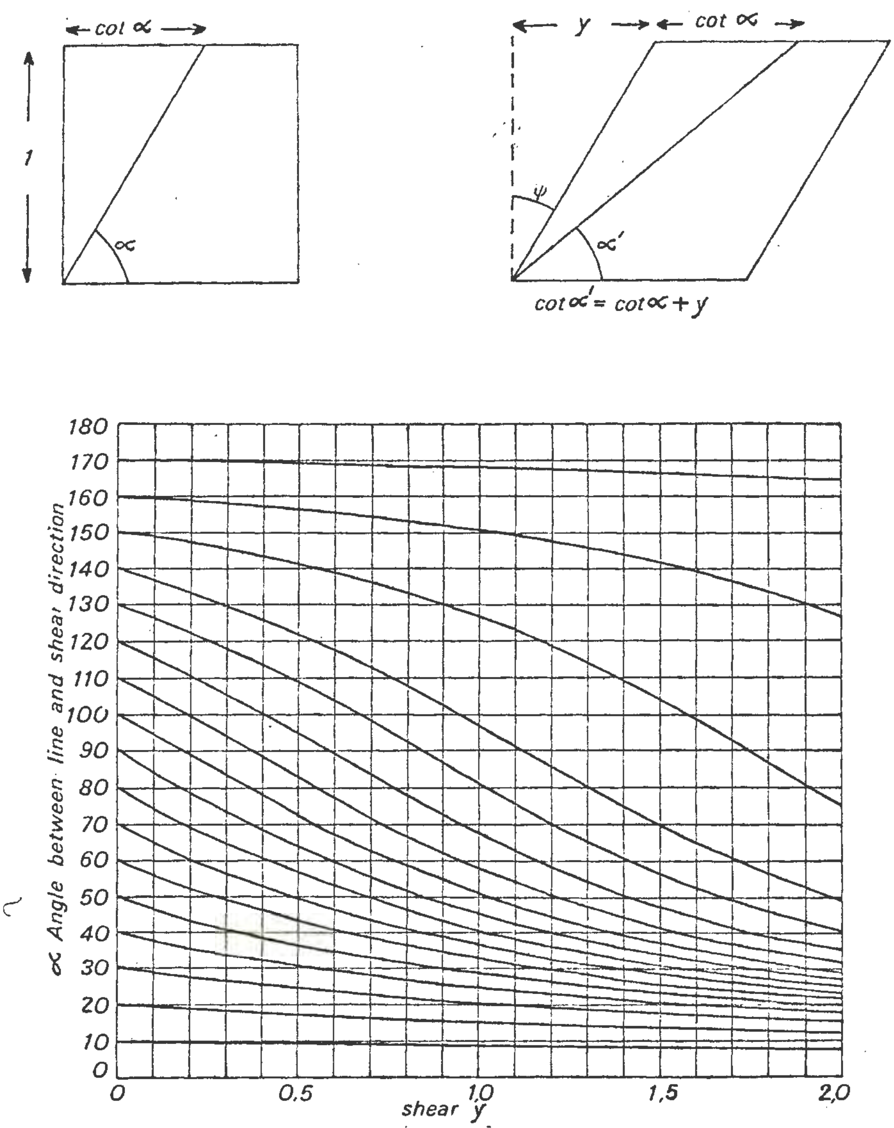


Figure 60. Change in orientation of a line making an angle α with the shear direction during increasing simple shear deformation. From Ramsay (1967, Figs. 3-23 and 3-24).

An illustration of this tightening effect is given in the case of folds having the same axial plane orientation as the D_5 structures, i.e. parallel to the b axis of simple shear and with their axes contained in the ab plane. As noted above the initial orientation of the axial planes was 45° to the shear direction and with this information the change in dihedral angle can be adequately plotted on a graph as a function of changing shear strain values (Fig. 62). The insets of Fig. 62 show the change in shape and orientation of the fold and the strain ellipse for various shear strain values.

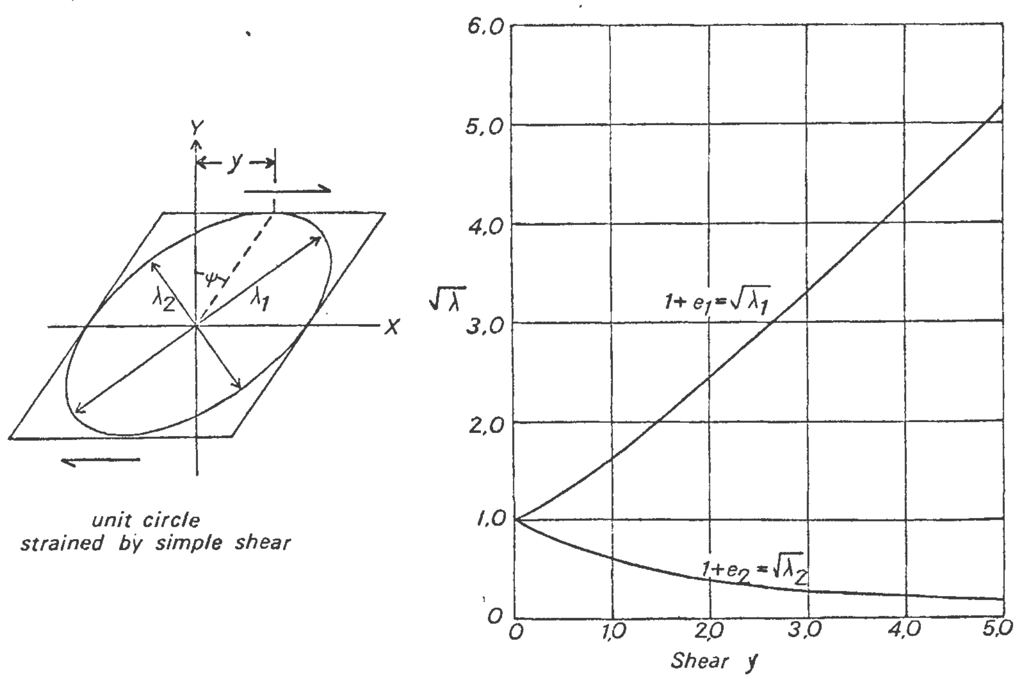


Figure 61. The ratio of the principal strain axes expressed as a function of shear strain during simple shear deformation. From Ramsay (1967, Figs. 3-20 and 3-21).

Several features of the Pofadder ZAHNCAFS D_5 structures show that they possibly formed in a zone of simple shear. These include the fold orientation the progressive change in fold orientation as well as the apparent tightening of these folds adjacent to the Lineament. Several features still remain unexplained and these are:

(i) absence of synforms between antiforms in some cases.

(ii) the doubly plunging effect of the antiforms.

(iii) the presence of 'cross folds' (D_6 folds trending parallel to the Lineament and deforming the D_5 structures).

(iv) absence of D_5 folds south of the Lineament

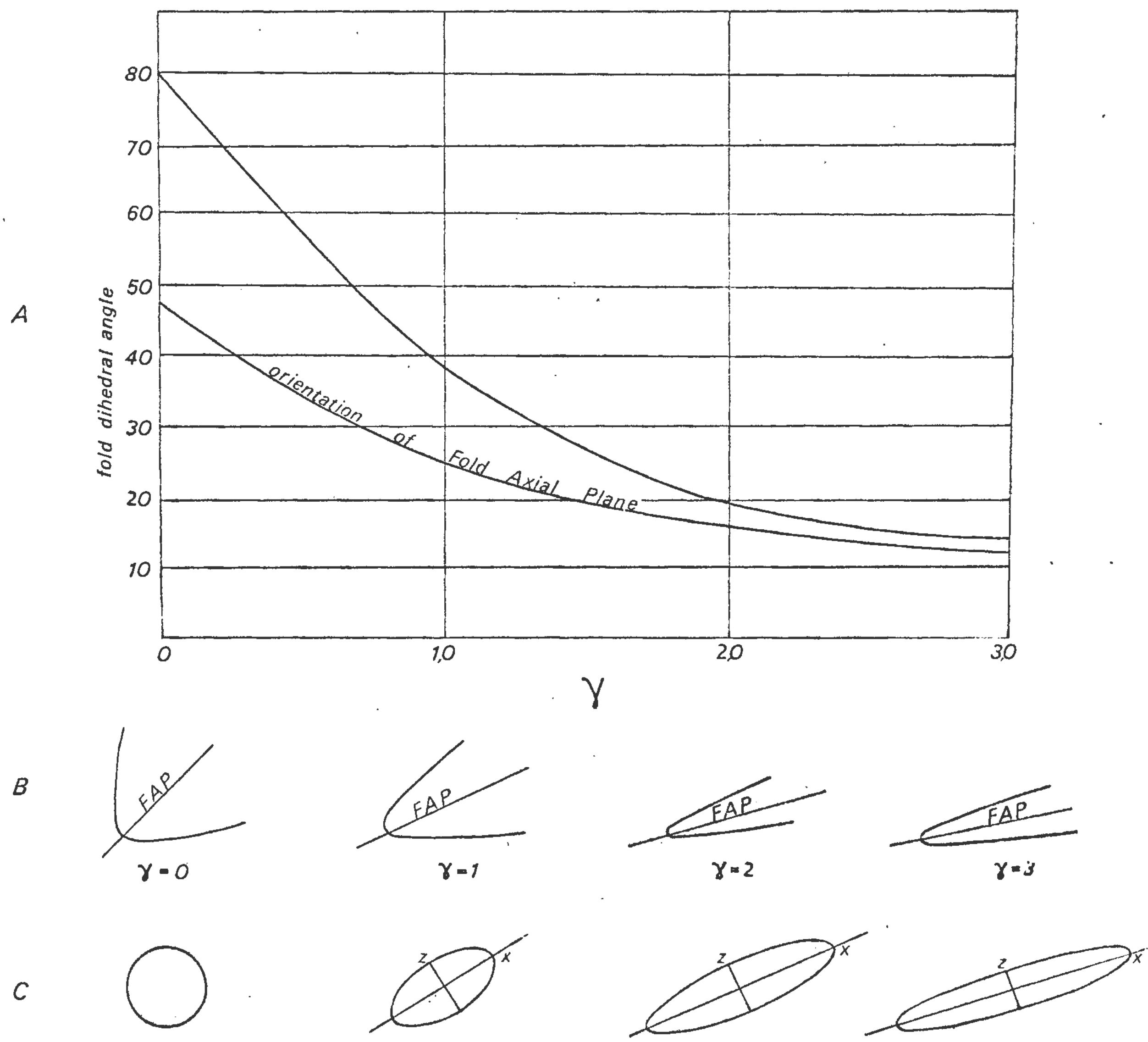


Figure 62. A. Variation of fold dihedral angle as a function of shear strain in a plane perpendicular to the shear plane and containing the shear direction. Initial orientation of fold axial plane 45° to the shear direction and initial dihedral angle 80°. Plunge of fold axis parallel to b axis of simple shear.

B. Change in shape and orientation of folds with given increments of shear strain.

C. Change in shape and orientation of the strain ellipse at the same increments of shear strain.

Points (i), (ii) and (iii) are part of a refolding effect and they indicate a compressive strain (pure shear) acting perpendicular to the boundary of the ZAHNCAFS. In the paper by Ramsay and Graham (1970), and echoed by successive workers since then, the presence of a pure shear component in the formation of shear zones was discounted. The general consensus is summarized in the following statement by Escher and Watterson (1974, p.223): 'Boundaries between undeformed rocks subjected to pure shear deformation must inevitably result in discontinuities because the initial boundary plane (which is the principal plane of the strain ellipsoid) changes in area in deformed rocks but not in the adjacent undeformed rocks'.

While this statement is perfectly true for rocks deformed by homogeneous pure shear it is not true for rock deformed by heterogeneous pure shear and the failure to recognise this point may have led to some inaccurate extrapolations of the Ramsay-Graham model. This is demonstrated in Fig. 63 which show that identical 'shear zones' in rocks containing a previous anisotropy can be formed by entirely different processes.

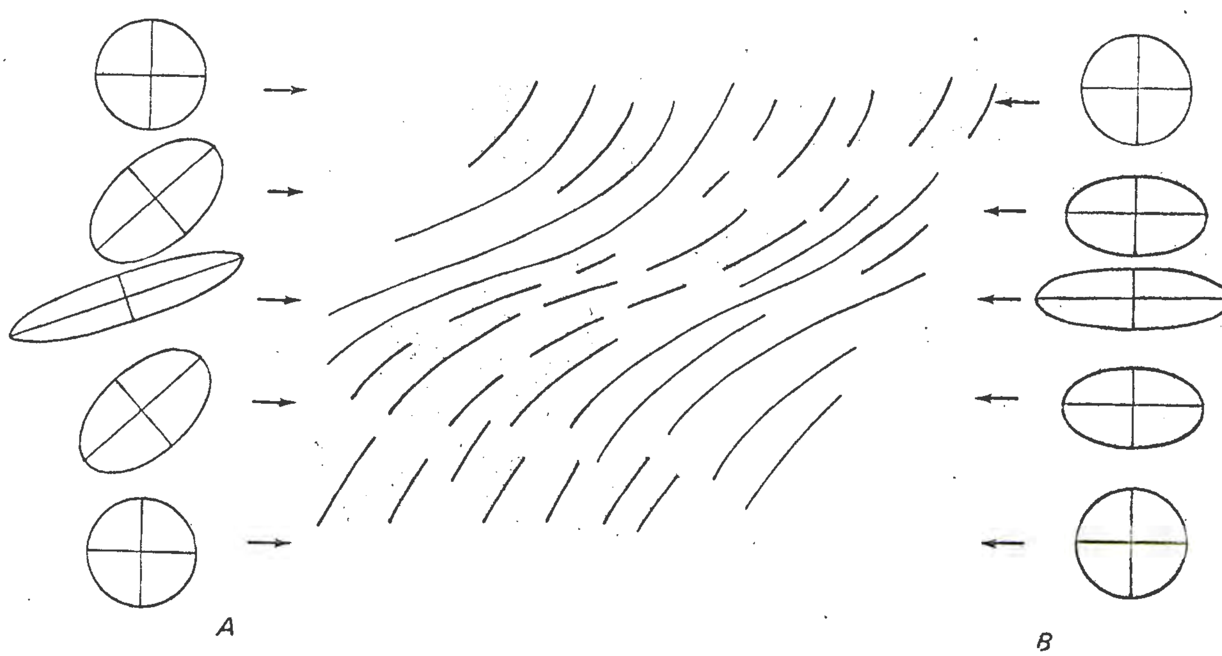


Figure 63. The formation of a 'shear zone' by progressive deformation involving either heterogeneous simple shear (A) or heterogeneous pure shear (B).

In the model of Ramsay and Graham the idea of a pure shear component was relegated on the grounds that the rocks outside the shear zone in question were undeformed and the geometric form of structures within the zone were very close to those predicted by simple shear. In their model it is very difficult to envisage the fabric being produced by heterogeneous pure shear alone, but can components of pure shear strain be entirely ruled out? If deformation resulted in a combination of pure and simple shear and if the strain increments were infinitely small the main geometric criterion of simple shear will still exist, i.e. at the zone boundary the orientation of the foliation would still be at an angle of 45° to the shear direction.

The fact that simple shear is probably responsible for shear zones in isotropic rocks is not disputed since pure shear alone can not explain the geometry and it is exceedingly difficult to envisage a mechanism by which pure shear deformation could become so localised. But possible pure shear components must remain a very real problem since a large pure shear strain can be accommodated in the model without becoming detected. Therefore, studies estimating the amount of crustal shortening (Escher and Watterson 1974) or displacement (Beach 1974; Coward et al. 1973; Bak et al. 1975b) across shear zones may be inaccurate

If one considers a pure shear component with the Z-axis acting perpendicular to the ab plane of simple shear then points (i), (ii) and (iii) listed above are no longer major obstacles. The D_6 fold axial planes are therefore parallel to the XY plane of the pure shear strain ellipsoid and the doubly plunging nature of the D_5 folds and the 'absence' of synforms then reflects the type of interference pattern produced by the D_5 and D_6 structures (Fig. 25, type-H).

Seen on a regional scale there appears to be a progressive increase in the size of D_5 folds when traced in an easterly direction. The Falls antiform, the most easterly D_5 structure whose size is known, is nearly 40 km long and 18 km in wavelength. This variability in fold size is difficult to explain in terms of the Ramberg/Biot theories on folding (Ramsay 1967, pp. 372-382) since these demand that the wavelength of buckling instabilities should be uniform.

Price (1975, p.572) has pointed out that the Ramberg/Biot theories may not be valid in all cases. He has shown that many fold trains develop serially rather than incrementally and the progressive development of these structures results in folds much larger in wavelength and amplitude at one end of the train than the other. The writer believes that this interpretation may be applicable to the D_5 folds in the Onseepkans area. It is noteworthy that the high-strain end of the D_5 fold train is terminated by the Naros complex, a large, rigid, intrusive massif extending across the Orange River east of Onseepkans.

3. *Strain analysis in the Pofadder ZAHNCAFS*

The change in shear strain values across the ZAHNCAFS can be estimated by the reorientation of the foliation as described by Ramsay and Graham (1970, p.802), and from these figures the displacement across the zone may be computed.

For isotropic rocks in which a foliation is produced the limits of a

ZAHNCAFS are easily recognised. However, it is not possible in anisotropic rocks since the limits of the zone cannot be identified with precision and the original orientation of the foliation outside the zone is not known. The ideal method in these cases (failing the presence of an identifiable ZAHNCAFS foliation) would be to plot the change in orientation of the axial planes of ZAHNCAFS folds. In the Onseepkans area very few minor structures associated with these D_5 folds have been recognised and, therefore, no direct measurements were possible in the field. Estimates of fold axial plane orientation taken from the map are not considered reliable enough.

Since one of the aims of studying a ZAHNCAFS of this nature is to compute the displacement, it is considered essential to accept a conservative figure for the width of the zone in order to avoid an overestimate. In practice, this determination proved not to be critical as the bulk of the displacement figure is given by the high shear strain values which are only found in the vicinity of the Lineament. The method proposed by Ramsay and Graham (1970) provides a minimum displacement figure since foliation directions parallel to the ab simple shear plane (the mylonite belt) cannot be used in a computation.

Most recent studies (Coward et al. 1973; Escher and Watterson, 1974; Bak et al. 1975a, 1975b) have recorded maximum shear strains of 6 or less which corresponds to a foliation that is oriented 5° to the shear direction. This is a low shear strain figure (c.f. Ramsay and Graham 1970, Fig. 6) and probably reflects the small scale mapping involved and the reliability of the field data. In traverses across the Pofadder ZAHNCAFS, foliation readings were made to the nearest 1° and the final figure chosen for the computation was an average of several readings taken at each locality. This enabled the writer to define a maximum shear strain figure of approximately 18. The choice of a suitable traverse line across the Pofadder ZAHNCAFS presented certain problems. This is due to the fact that the change in foliation orientation is not only due to the presence of the zone but is also influenced by the folds on either side of the Lineament. The line finally chosen is shown in Fig. 56 and avoids fold hinge zones which would cause considerable fluctuation in shear strain values. The choice of the correct shear direction is also a potential source of error and in this case the shear direction was arrived at by averaging the mylonite foliation orientation along the traverse line which gave a shear direction of 107°. For the purposes of the computation the northern limit of the ZAHNCAFS was taken slightly inside the axial trace of the Shining Sister antiform where there is still an obvious relationship between the orientation of the foliation and the Lineament. It is emphasized that this is thought to be well inside the proposed limits of ZAHNCAFS development. The strike of the foliation at the limit of the zone on the traverse line is 65°. The southern limit was taken at the Orange River since no reliable data were available south of the study area. However, the air photo interpretation suggests that the immediate zone of reorientation south of the Lineament ends a short distance south of the river. The shear strain across the traverse line is given by the formula:

$$\cot \alpha^1 = \cot \alpha + \gamma \quad \text{(Ramsay 1967, p.88)}$$

Where α = angle between the shear direction and the foliation at the margin of the zone and α^1 represents subsequent foliation orientations as the core of the zone is approached.

$$\therefore \gamma = \cot \alpha^1 - \cot \alpha$$

The displacement across thezone is found by graphing the changing shear strain values and integrating the area under the curve (Ramsay and Graham 1970). This has been done for the Pofadder ZAHNCAFS and the result is shown in Fig. 64. This graph is a best fit 6th order polynomial regression strain distribution curve (computer programm BMDO-5-R Health Sciences Computing Facility, University California, Los Angeles, 1965), which gives a displacement figure of 100 km.

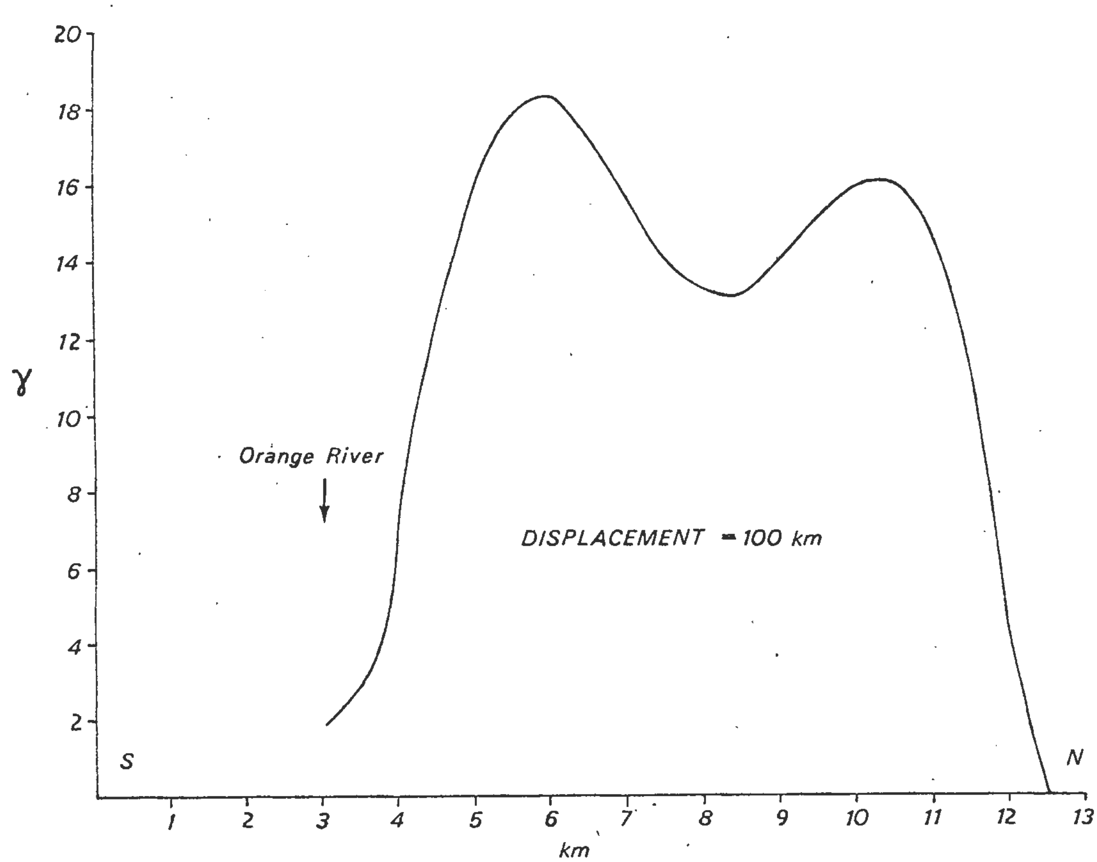


Figure 64. Strain distribution curve across the Pofadder ZAHNCAFS assuming a formation by heterogeneous simple shear. Trend of zone 107°, initial orientation of the foliation at zone boundary 65°, $\alpha = 42°$.

It has been suggested above that the Pofadder ZAHNCAFS did not result from heterogeneous simple shear alone but also involved a pure shear component. It is quite conceivable that this component was a post-ZAHNCAFS flattening but it is more likely that it acted incrementally. Obviously any pure shear component will have the effect of reducing the apparent displacement computed by the above method. The actual amount may be estimated for any given pure shear component where X = a, Y = b and Z = c, since the angle which the foliation makes with the shear direction before and after deformation is given by :

$$\tan \theta^1 = \tan \theta \left(\frac{\lambda_1}{\lambda_2}\right)^{\frac{1}{2}} \quad \text{(Ramsay 1967, p.67)}$$

Where θ^1 is the angle which the foliation makes with the shear direction after a combined pure shear/simple shear deformation, and θ is the angle resulting from simple shear only.

The actual lengths of the pure shear strain ellipse axes need not be known as the formula involves only the ratio of these parameters.

To test the effects of incremental pure shear components on the displacement figure given above two further strain distribution curves were plotted for 20% and 40% pure shear strains corresponding to $\lambda_1:\lambda_2$ of 1:1,6 and 1:2,8. The strain component considered incrementally varies from a maximum value in the core of the zone to zero at the boundary (to facilitate the computation only the pure shear component was added to the simple shear model as a post-simple shear heterogeneous pure shear flattening effect). The above pure shear strain components are considered realistic and approximately compatible with the dihedral angle of the D_6 folds forming the dome and basin structures at the margin of the ZAHNCAFS. These curves are reproduced in Fig. 65 and give displacements of 90 km and 85 km respectively. Pure shear strains of the above amount, if considered as a post-ZAHNCAFS effect and not incrementally, will result in slightly greater reductions in apparent displacement. A reduction of 14% in apparent displacement results from a 40% incremental homogeneous pure shear component in the formation of the ZAHNCAFS which suggests that the displacement estimates based on a simple shear model alone provide solutions of the right order of magnitude.

It has been assumed here that the axes of the pure shear strain ellipsoid are roughly coincident with those of the simple shear ellipsoid (Fig. 66A). There is no fundamental reason why this should be so and it would be more compatible with the suggested model to have the pure shear X-direction parallel to b and thus sub-parallel to the Y-axis of the simple shear ellipsoid (Fig. 66B). This would then produce a maximum extension direction in the surrounding rocks perpendicular to the axial trend of D_6 folds. This consideration would still further reduce the disparity in displacement produced by a pure shear component acting in the ZAHNCAFS.

The incremental pure shear component has not only accentuated the finite strain heterogenity but has also produced fold interference patterns varying from type-H crescent shaped folds near the core to type-B dome (Fig. 25) dome and basin structures at the boundary.

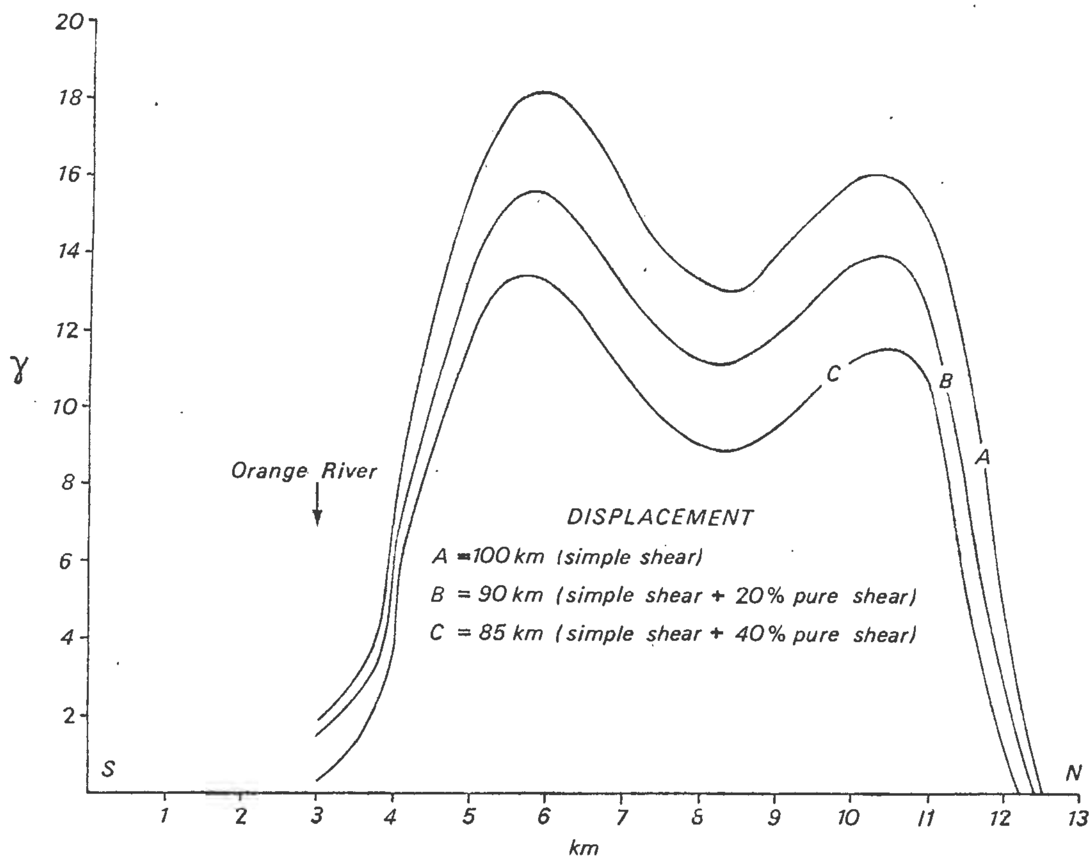


Figure 65. Best-fit strain distribution curves across the Pofadder ZAHNCAFS assuming a combination of heterogeneous simple shear and incremental homogeneous pure shear components. Curve from Fig. 64 included for comparison.

4. *Relationship between the mylonite fabrics and the zone of re-orientation.*

In Section IVA3 it was pointed out that the structures and fabric elements in the mylonites tended to have the same orientation as the structures in the surrounding gneisses (c.f. Plate 41). It is now suggested that this mirror image effect is due to the strain pattern in the core of the XAHNCAFS that existed during the period of refoliation.

It can be seen in Fig. 61 that even for modest amounts of shear strain ($\gamma = 3$) the orientation of the X-direction of the simple shear strain ellipsoid is sub-parallel to the ab simple shear plane. For further strain increments the greatest direction of shortening in the rock will be virtually perpendicular

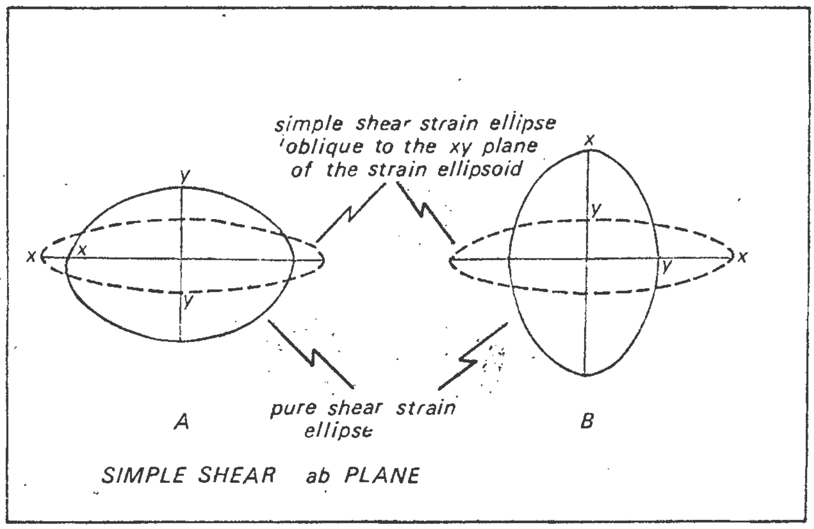


Figure 66. Schematic representation of two possible combinations of pure and simple shear that may be applicable to the Pofadder ZAHNCAFS.

to the ab plane. This shortening will be reinforced by the pure shear component regardless of the orientation of the X-direction and assuming only the Z-direction to be perpendicular to the ab plane.

In the bc plane of simple shear the Z:Y axial ratio of the simple shear ellipse at γ = 18 is 1:3, and the Z:X ratio of the pure shear strain ellipse corresponding to a 40% strain is 1:2,8 or, in the situation where the X-direction is parallel to b,Z:Y is 1:1,7. The shape of the combined pure shear zone/simple shear strain ellipse can be computed by simple matrix operation (Ramsay 1967, p.322-326) and Fig. 67 illustrates how the final shape depends on the orientation of the pure shear component. The two ellipses have ratios of 1:8,7 or 1:5,3.

If the pure shear component in a case such as Fig.66B is sufficiently large this may change the Flinn plot (c.f. Fig. 44) from a type 1<k<∞ to a type 1>k>0, that is, to one of apparent flattening. Thus it can be seen that while the overall strain pattern in the ZAHNCAFS is one of hetereogeneous simple shear strain and 'stretching' (Escher and Watterson 1974) the core of one of these zones may be characterized by strong flattening strains.

The rotation and tightening of folds adjacent to the Lineament has been described above but Ramsay (1962, 1967, pp.411-415) has pointed out that there is a limit at which shortening in layered rocks can no longer be accommodated by folding and beyond which the superimposed homogeneous strain produces flattened folds. The flattened folds in the core of the ZAHNCAFS which have their axial planes parallel to the ab plane of simple shear will result in the fold hinge areas moving away from one another in the ab plane (Fig.68).

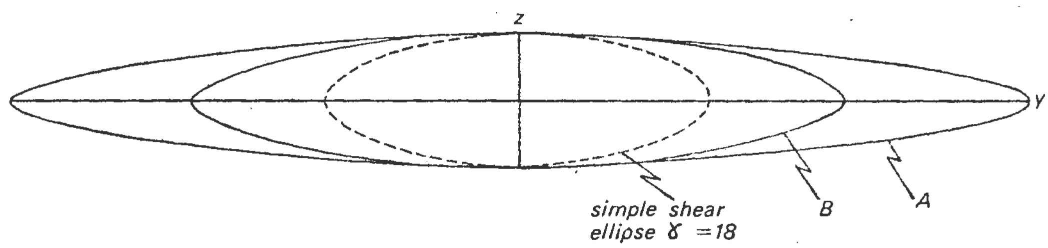


Figure 67. Shape of the YZ plane through the combined pure shear/simple shear strain ellipsoid parallel to the bc plane of simple shear and assuming a 40% pure shear component and γ = 18. A. Xps=Yss, Yps=Xss, Zps=Yss. B. Xps=Xss, Yps=Yss, Zps=Zss.

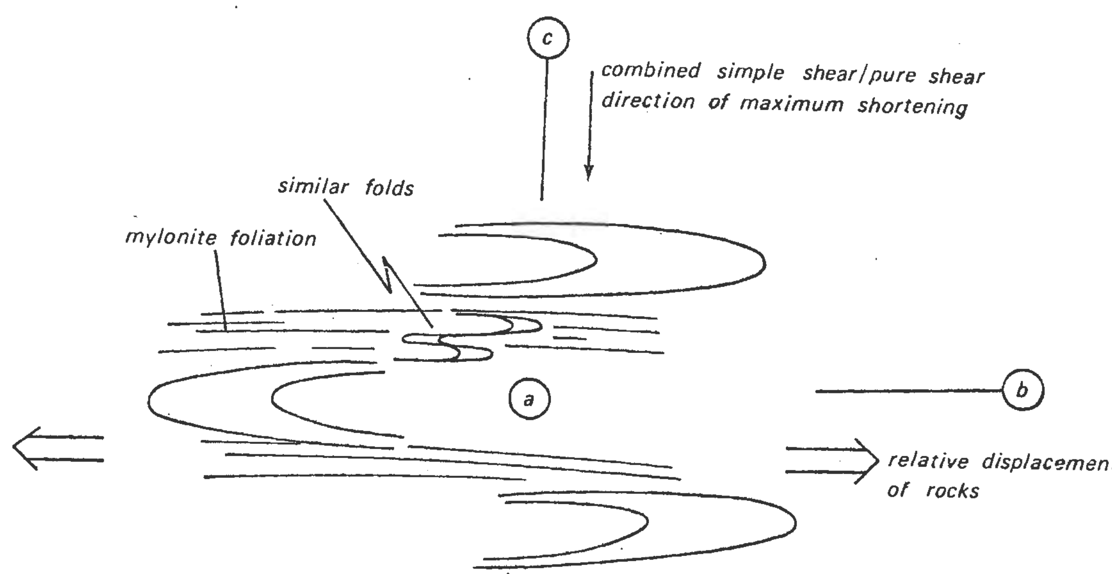


Figure 68. Section parallel to the bc plane of simple shear in the core of the Pofadder XAHNCAFS showing the relationship of the mylonite foliation to the flattened folds.

It is suggested that refoliation of the core rocks during this stage of ZAHNCAFS formation has produced mylonites oriented parallel to the finite strain ellipsoid XY plane, the fold axial planes and, because of the high shear strain involved, parallel to the ab plane. Folds in the mylonites are probably of similar type (Plate 40) and have the same orientation as the flattening structures. In the hinge zone of the Mission antiform and immediately adjacent to the Lineament a fold axial plane cleavage is unquestionably defined by this mylonite fabric which contains a lineation produced by sillimanite needles (thin section DT 1093, Plate 46). Away from the core of the zone the mylonite foliation appears to be confined to the more incompetent hornblende gneisses that form envelopes around the D_5 fold cores consisting of quartzo-feldspathic gneiss. This produces small bifurcations from the main mylonite belt.

5. *The development of the structural pattern south of the Pofadder Lineament and pressure shadow effects*

The absence of D_5 and D_6 folds south of the Lineament is strikingly apparent in the Onseepkans area (Fig. 56) and the change in structural style has also been noted in the Warmbad area by Beukes (1973). This distinction is illustrated by the orientation diagrams for the southern block (Fig. 57, zones 5A, 5B) which shows that the zone with fabrics parallel to the shear direction is much wider here than in the northern block where the ZAHNCAFS folds have caused a more abrupt deflection of the earlier fabrics.

The model for the development of the ZAHNCAFS outlined in the preceding pages predicts the presence of D_5 and D_6 structures in this area and their absence requires some modification to the hypothesis. Several contributory factors have affected the style of deformation in the southern block. It is known, for example, that P-T conditions here were lower than in the area north of the Lineament and the effect of varying P-T parameters on deformation is well documented (Griggs et al. 1960; Heard 1960; Biot 1961, pp.1607-1611). The writer believes, however, that the simplest explanation is found in the relationship of the southern block to the neighbouring Vioolsdrif complex.

When Fig. 56 is examined it can be seen that in an area between the Vioolsdrif complex and Tantalite Valley there is no reorientation of folds and foliation as the Pofadder Lineament is approached. Only east of Tantalite Valley is this shearing effect noticeable. This is a classic example of a pressure or strain shadow which is produced in the low-strain area adjacent to a rigid incompressible body perpendicular to the maximum shortening direction.

The deformation of materials containing rigid isotropic bodies has been investigated theoretically and experimentally by Savin (1966) and Strömgard (1973). The extent of the pressure shadow depends on the shape of the rigid body and ratio of applied stresses and an example of the role played by the latter is shown in Fig.69. It is suggested, rather speculatively, that the Pofadder ZAHNCAFS could represent the interface between compressive and tensile stresses in this example as shown in Fig.70.

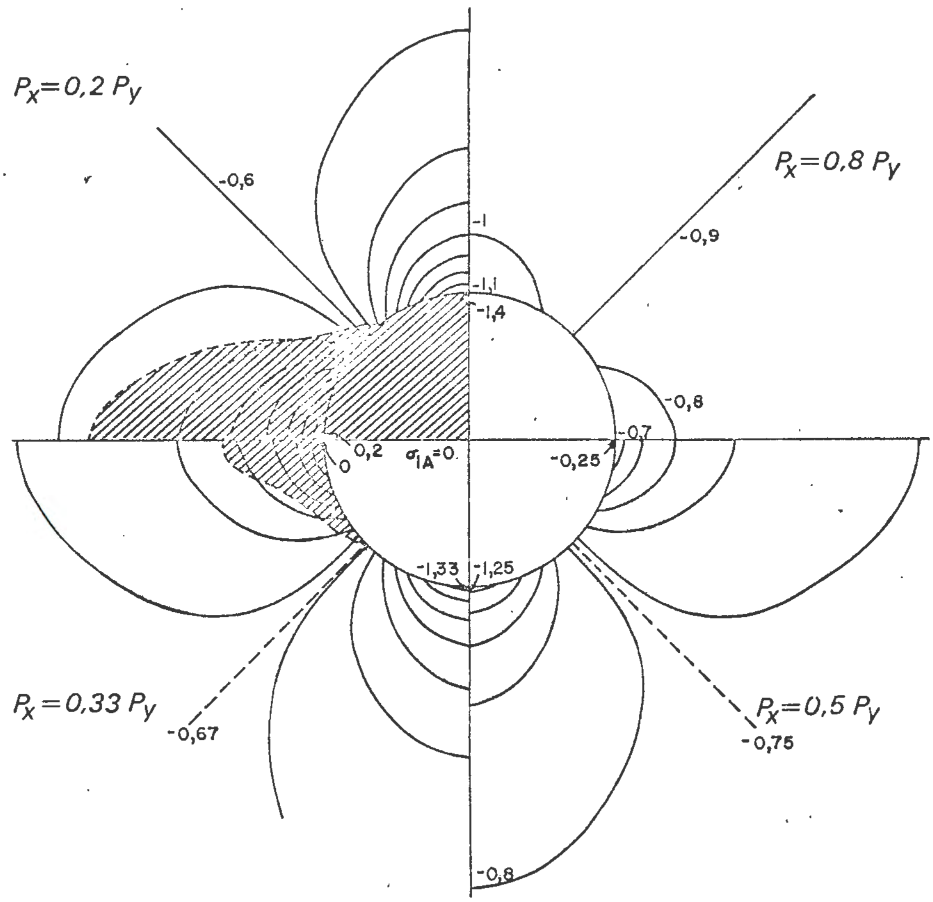


Figure 69. Stress distribution inside and outside a rigid incompressible body for different values of applied stresses. Px = horizontal stress, Py = vertical stress. Shaded areas represent areas of tensile stress. (Strömgard 1973).

The pressure shadow region in the Pofadder ZAHNCAFS obviously encloses the area from the Vioolsdrif complex to Tantalite Valley but the distinction in structural style across the Lineament persists well to the east of Tantalite Valley and suggests that the pressure shadow effect may be more extensive. Perhaps the strain shadow hypothesis can explain the absence of D_5 and D_6 folds in the southern block on the grounds that this area has been isolated from the pure shear component in the ZAHNCAFS formation. The 'extension' of the strain shadow between Tantalite Valley and Pofadder therefore illustrates ZAHNCAFS formation by the classic example of simple shear only.

Although the strike slip origin for the Pofadder ZAHNCAFS is well proven on structural grounds the apparent displacement is not compatible with lithological observations. In Chapter III it was shown that a very marked change in metamorphic parageneses takes place across the Pofadder Lineament defined by high-grade and granulite-grade assemblages in the northern block and medium-grade assemblages in the south. It was proposed that vertical displacement

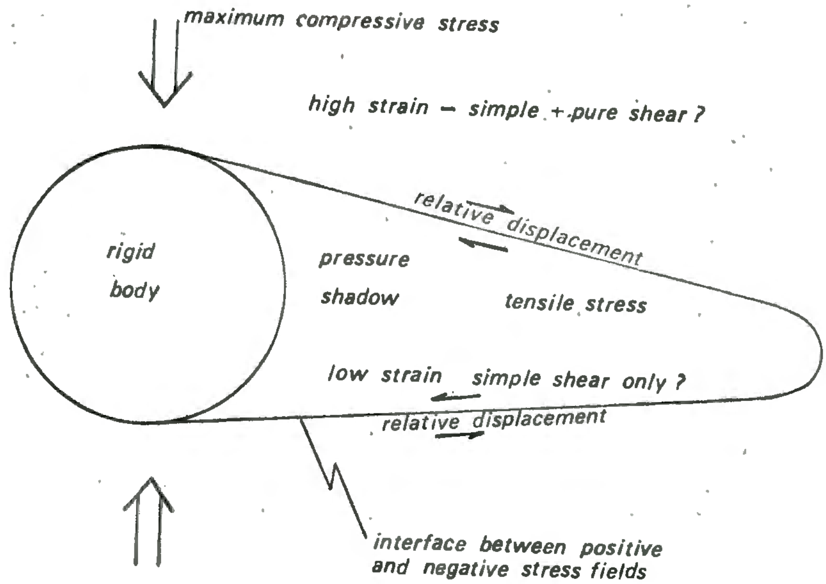


Figure 70. Distribution of stress and strain around a rigid body showing hypothetical displacements along interfaces between positive and negative stress fields.

along the Lineament could account for this; in other words, the southern block represents a higher crustal segment that has been downthrown relative to the northern block.

The strike slip displacement may reflect an early stage in the development of the ZAHNCAFS up to the formation of the mylonite belt. It has been shown in section IVB*1* that mylonites, being weak rocks, will localise any subsequent stress and it is therefore acceptable on theoretical grounds to suggest that the vertical component of movement represents a closing stage in the development of the ZAHNCAFS. A similar disparity in displacement evidence is mentioned by Sutton and Watson (1962, p.96) along many of the small shear zones represented by the Laxford Front in the Scottish Highlands. The reorientation of a pre-existing foliation in most of these zones indicates vertical movement but in some cases a horizontal displacement of up to 0,5 km was shown by dyke displacement.

6. *Regional considerations*

The area east of the Vioolsdrif complex and south of the Pofadder Lineament in which medium-grade metamorphites are preserved appears to be limited in extent because, further to the south, norites and granulite-grade assemblages are reported in the Pofadder-Aggeneys region by Joubert (1974b). The southern limit of the norite-free zone roughly coincides with an east-west trending discontinuous shear zone (Joubert 1974a, Fig. 1, and maps in preparation) south of

Pofadder and north of Aggeneys. It is possible, therefore, that the medium-grade crustal segment is bounded by mylonite rocks both in the north and in the south. The area west of Aggeneys has not been studied but mylonite belts mapped by Joubert may extend some distance further towards the coast, south of the Vioolsdrif complex. East-west trending mylonite belts have been described in the northwestern Cape Province at this lattitude by Gevers et al. (1937) and in the eastern Richtersveld by De Villiers and Söhnge (1959, p.112) and mylonites with a similar trend have been noted in the area north of Steinkopf by Bertrand (pers. comm.).

Significantly, a series of northeast trending en-echelon doubly-plunging folds have been mapped south of the mylonite belt in the Pofadder-Aggeneys area (Fig. 71) and therefore they present a similar situation to the Pofadder ZAHNCAFS which bounds the Vioolsdrif complex in the north. Fig.71 should be compared with Fig. 70 showing the strain shadow patterns around a rigid body. It should be noted however, that the displacement directions indicated by these folds is not the same as that for the Pofadder ZAHNCAFS.

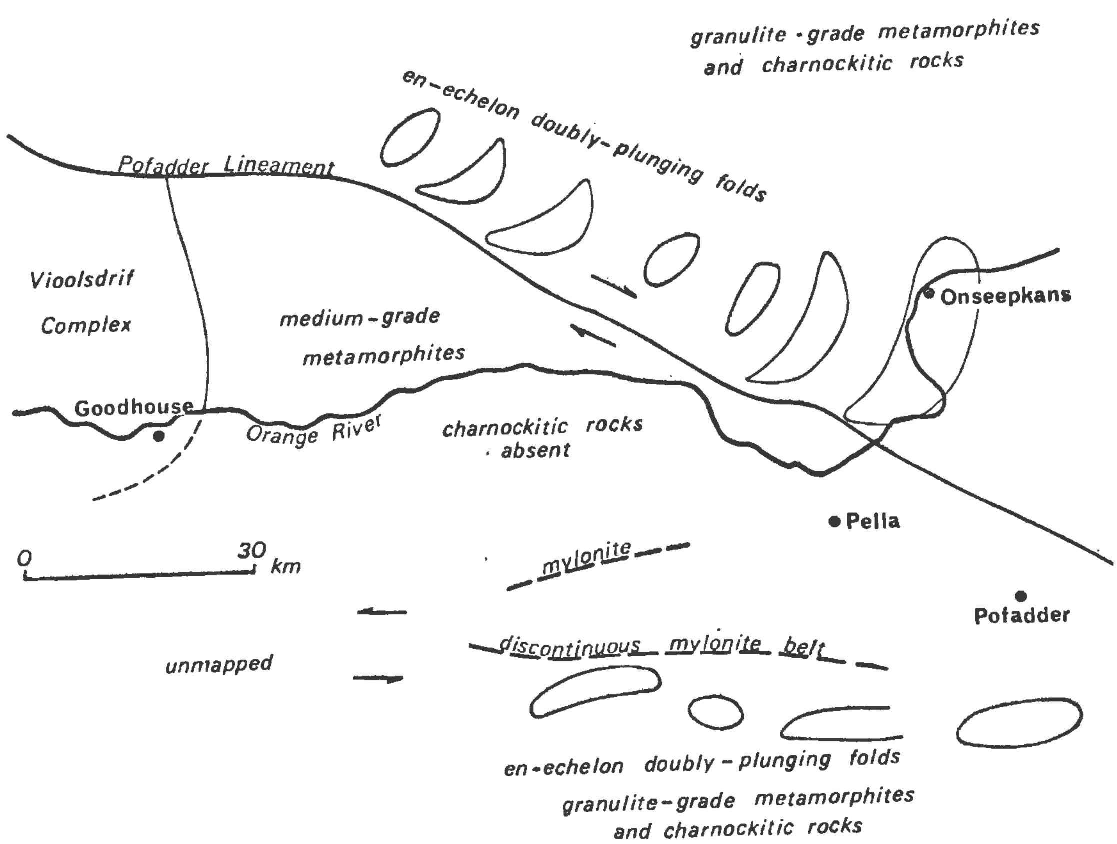


Figure 71. Sketch map of the Goodhouse-Onseepkans area showing the disposition of mylonite belts and associated en-echelon doubly-plunging folds in relationship to the Vioolsdrif complex.

The distinction between the medium-grade and high-grade metamorphic assemblages across the Pofadder Lineament has also been noted by Jackson (1976) 300 km to the northwest in the Aus area (see Fig. 24) and this would mean that the model tentatively proposed here may be applicable to a large portion of the Namaqua belt. This could imply that the Vioolsdrif complex and its envelope of medium-grade metamorphites is bounded by shear zones and is a large graben structure.

This interpretation also gains some support from radiometric data. All granitoids of the Vioolsdrif complex yield ages of 1750-1900 Ma (Kröner 1974b). A minimum age of ∿1755 Ma has been obtained from a mafic gneiss at Lüderitz (Kröner 1975) and a minimum age of ∿1575 Ma was obtained from a mafic gneiss south of the Pofadder Lineament near Pofadder (Kröner 1974b). All ages obtained on Namaqua belt rocks outside the 'graben', including the Springbok area to the southwest (Clifford et al. 1975), the Aus, Ai-Ais and Onseepkans areas to the north and the Upington area to the east (Kröner 1974b, 1975; Nicolaysen and Burger 1965) all give ages varying between 900-1200 Ma with only one exception from the Onseepkans area. No details about methods etc. are taken into account here and the interpretation of such variations in preliminary age dates is controversial, but the 1750-1900 Ma ages may reflect the earlier closing of the system in the higher level rocks now forming the graben (York and Farquahr 1974, pp. 100-108). The various radiometric, metamorphic and structural attributes of the 'graben' are shown in Fig. 72 and the radiometric data for rocks in the Onseepkans area are shown in Appendix 1.

This conjectural model for the geology of the Namaqua belt and the Vioolsdrif complex differs from recent interpretations by Blignault (1974) and Blignault et al. (1974). In Blignault's model the 'grey gneisses' in the Namaqua belt are interpreted as having formed from the older, unfoliated Vioolsdrif complex by progressive metamorphism and associated deformation. The non-penetrative mylonite foliation seen in the Vioolsdrif complex, for example, is equated with the first tectonic fabric (s_1) developed in the mobile belt.

Blignault's hypothesis is at variance with recent studies by Bertrand (1976) in the Henkries-Goodhouse area. There the Vioolsdrif Granite is described as intruding into foliated and migmatised 'grey gneisses' with subsequent metamorphism locally producing strongly foliated and migmatised gneisses from the intrusives. The low-grade mylonite foliation is stated to be superimposed on these earlier fabrics and this would reinforce observations by the writer from the Warmbad area that the mylonite foliation in the granite is associated with a similar foliation in the Pofadder Lineament, i.e. it is the latest fabric developed in the rocks and not the earliest as proposed by Blignault. The intrusion of the Vioolsdrif Granitoids into the 'grey gneisses' has also been recently noted in the Ai-Ais area (Kröner pers. comm.).

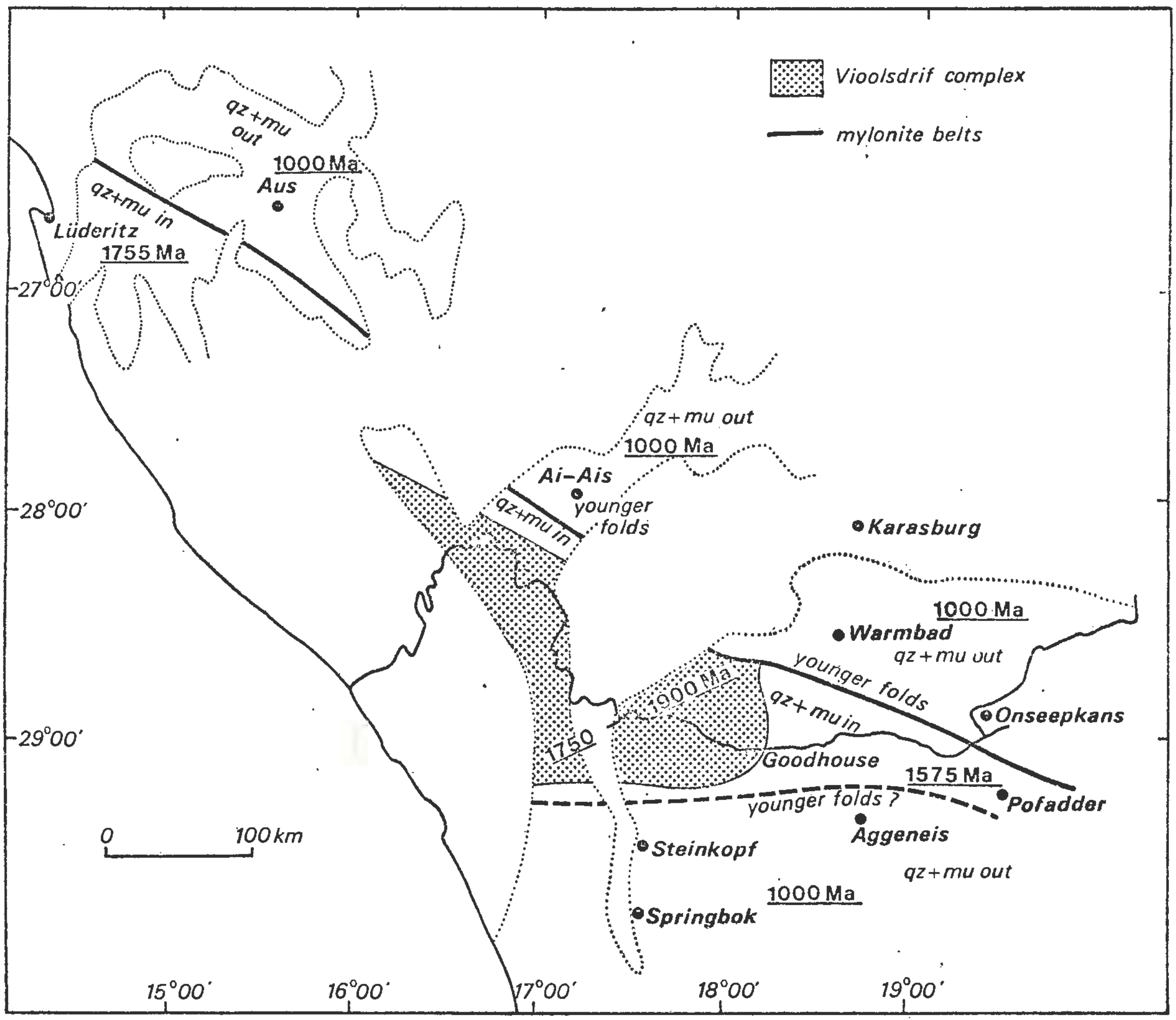


Figure 71. Outline of the 'Richtersveld graben' showing the distribution of radiometric age dates (abbreviated to 1000 Ma event outside the graben) metamorphic assemblages and structures. Data from Gevers et al. 1937; De Villiers and Söhnge 1959; Beukes 1973; Joubert 1974b; Blignault et al. 1974; Jackson 1976 and Kröner 1974b, 1975.

CHAPTER VI

MESOSCOPIC FOLDS AND FINITE STRAIN ESTIMATES

This chapter is concerned with the mesoscopic folds which are found throughout the Onseepkans area. Quantitative techniques are used in the study of these folds in order to determine the rheological properties of the rocks during deformation and the fold-forming mechanisms. Further techniques are employed to estimate the total finite strain suffered by the folded rocks. Considering the size of the area mapped this study cannot be regarded as an exhaustive analysis of mesoscopic folds or finite strain states and the limitations are detailed in section VIA3

The following section outlines the theoretical background to the study of folds and will serve as a basis for the succeeding analysis.

A. INTRODUCTION

1. *Fold development.*

During the last 15 years the study of folding in naturally deformed rocks has benefited enormously from a steadily growing amount of experimental and mathematical work (Biot 1961, 1964, 1965a, 1965b; Biot et al. 1961; Ramberg 1959, 1961a, 1961b, 1963a, 1963b, 1964; Chapple 1968, 1969; Sherwin and Chapple 1968; Dieterich and Carter 1969; Ghosh 1968; Ghosh and Ramberg 1968; Price 1967; Ramsay 1967; Patersen and Weiss 1966; Cobbold et al. 1971; Handin et al. 1972; Hudleston 1973a, 1973b; Hudleston and Stephansson 1973; Currie et al. 1962; Johnson and Honea 1975; Donath 1971; Mathews et al. 1971; Gay and Weiss 1974; Hobbs 1971; Flinn 1962; Stabler 1968; Skjernaa 1975). Perhaps the most important conclusion reached in some of these studies is that folding is related to the thickness of the layer being deformed and the viscosity contrast between this layer and the surrounding material. The expression connecting these parameters is:

$$Wd = 2\pi t \sqrt[3]{\frac{\mu_1}{6\mu_2}}$$

(Biot 1961, p.1614; Ramberg 1961a, p.418)

where Wd = dominant fold wavelength
t = thickness of the layer
μ_1, μ_2 = viscosity of layer and medium respectively

Although this expression has subsequently been modified its applicability has been demonstrated in experimental work and in naturally deformed rocks (Sherwin and Chapple op. cit; Ramberg 1964; Biot et al. op. cit; Currie et al. op. cit; Hudleston 1973b, 1973c). The most significant modification to the theory resulted from the realisation that materials deform homogeneously prior to buckling (Sherwin and Chapple op. cit.), an effect that is now commonly referred to as 'layer-parallel shortening'. This enables the above expression to be rewritten as:

$$Wd = 2\pi t \sqrt[3]{\frac{\mu_1}{6\mu_2}\,\frac{1}{2}\,\frac{s-1}{s^2}}$$

(Hudleston 1973a, p.29)

where $s = \sqrt{\lambda_1/\lambda_2}$ of the homogeneous strain

A further limitation has been suggested by Chapple (1968, p.50). He has shown that the Ramberg/Biot theory only holds for folds with limb dips of less than 15°. Beyond this angle the development of the fold is related to its amplitude and the actual wavelength (W) is dependent on the ratio W/Wd and limb dip.

2. *States of strain in and around the folded layer.*

This aspect of folding theory has been reviewed by Ramsay (1967, pp.386-411, pp.415-421) and Ramberg (1961a, 1961b, 1963, 1964). Both authors have established that there are two dominant processes by which folds develop; flexural slip or flexural flow (concentric shearing strain) and tangential longitudinal strain (concentric longitudinal strain). The most obvious difference between these two processes is that, in the former, the sites of greatest strain are found on the fold limbs and, in the latter, greatest strain occurs in the fold hinge. Two further fold producing mechanisms have been identified. Ramsay (1967, pp.421-436) discussed the problem of similar folds in which layers behave passively and are transposed or sheared by a closely spaced set of planes at an angle to the layering. Ramberg (1963a, pp.6-10) has suggested an analogy between similar folds and what he terms 'bending folds' which are folds produced by compression perpendicular to the layering in anisotropic rocks. 'Shear' folding encounters problems in that the mechanism does not account for the periodicity of these structures and it has been further shown by Ramsay (1967, pp.411-415) that similar folds may be regarded as normal buckle folds which have suffered an infinite amount of post-fold homogeneous flattening (see also section VIB3).

If there is sufficient viscosity contrast between the folding layer and the surrounding material the latter will simply be pushed aside by the developing fold and be squeezed out of its core. Approximately one wavelength away from the layer the surrounding material will deform homogeneously without folding. In buckled multilayers the more incompetent (less viscous) material will be squeezed into the hinge zones and will resemble a flexural flow fold.

3. *Fold classification.*

The analysis presented in sections B and C is based on certain techniques which will now be outlined.

The application of folding theory to naturally deformed rocks is largely dependent on the fold classification system developed by Ramsay (1967, pp.359-372). This is the only quantitively based fold classification system in existence and the only one that allows folds to be compared unambiguously.

Basically the system involves computations made on fold profiles. The thickness of the layer ($t\alpha$) is measured between tangents drawn on the bounding surfaces for equal amounts of dip (α) and this is expressed as a ratio (t_1) of $t\alpha/t_0$ where t_0 is the thickness of the layer measured in the fold hinge (Fig. 72A). The changing shape of the fold can then be represented by a graph of t against α. If lines (isogons) are drawn across the fold connecting points formed by the tangents at equal dip (α) then an additional tool for fold description is provided. Using these methods Ramsay (1967, pp.386-396) has defined 3 main classes of folds reproduced here in Fig. 72B. Class 1 is further subdivided into folds that have strongly convergent dip isogons (1A) where $t_1 > 1$ those that have convergent dip isogons (1B) where $t_1 = 1$ (parallel folds) and those with weakly convergent dip isogons (1C) where $t_1 < 1$. In class 2 folds the dip isogons are parallel and these represent similar folds. Class 3 folds are those with divergent dip isogons.

In the present study these computations were made on fold profile photographs taken from specially selected areas. Altogether some 207 photographs of folds were taken. Not all of these could be used in the analysis, the most common handicap being the inability to define exactly the contact between two layers. It was found in some cases that, where the outcrop surface did not represent a fold profile, the oriented photograph was out of focus in places. This also resulted in the limitation of the number of photographs taken since outcrop surfaces that were at a high angle to the fold profile could not be used. All fold classifications published so far have been on profile planes, and the writer did not investigate the possibility of using other sections through the fold for these analyses.

In a fold train isogons can be rapidly drawn without involving errors resulting from poor choice of bounding surfaces by the use of incoherent image processing (Lasserre and Smith 1974). This technique is less suitable for the analysis of individual folds or quarter wavelength folds.

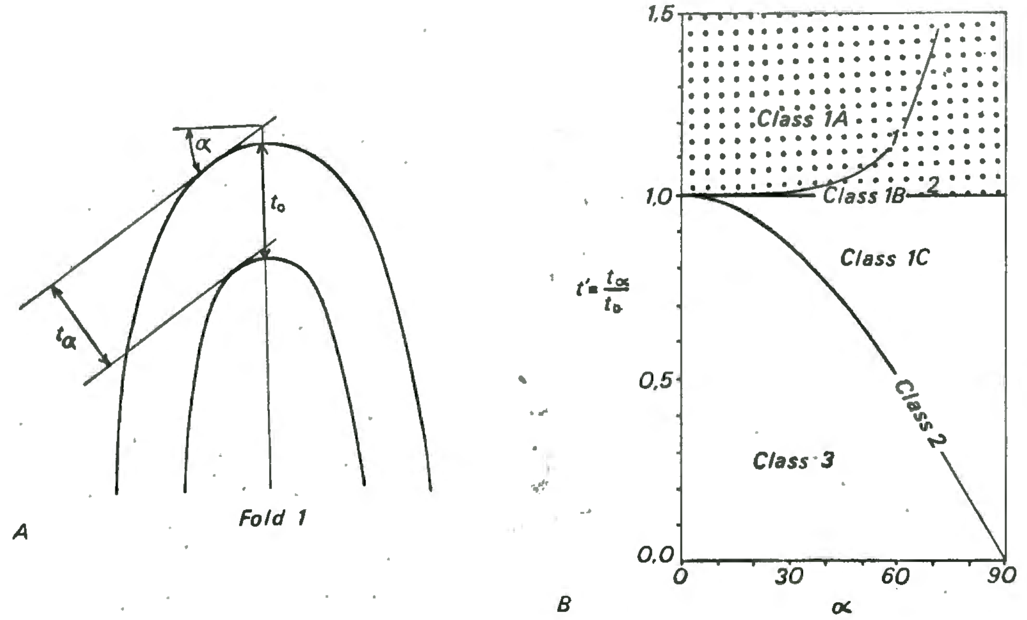


Figure 72 A. Fold profile showing relationship of layer thickness (t_α) to limb dip (α).

B. Fold classification system based on limb dip and layer thickness after Ramsay (1967, Figs. 7-18 and 7-25).

183 folds were classified in the present study, by far the majority of them are class 1C and the second most important group are class 3.

4. *Harmonic analysis.*

Not all folds are amenable to treatment by the above method. Folds defined by surfaces rather than by layers can be studied by Fourier analysis (Stabler 1968) in which folds are described by a sum of a number of sine and cosine harmonics of decreasing amplitude. In the present study 168 quarter-wavelength folds were analysed by this method using a digital computer (written by C.J. Hartnady 1974). The first (b_1), third (b_3) and fifth (b_5) harmonics were plotted; higher values have little significance since Hudleston (1973c) found that about 95% of the fold shape was given by the first harmonic and less than 1% by the fifth. The shape can therefore be adequately described in almost all cases by the first and third harmonics.

Computation was again made on fold profile photographs. The three invariant points on the fold were identified and a perpendicular was drawn from a tangent at the hinge to the line connecting the inflexion points (or to the point where a line parallel to the hinge tangent intersects one inflexion point in asymmetric folds). The quarter-wavelength is then divided according to the harmonic number required and measurements are made from this axis to the fold surface. The results can be plotted on a graph of b_1 against b_3 for comparison or as frequency histograms.

Since harmonic analysis describes attributed such as fold shape and 'tightness' it is equally applicable to many of the normal layer folds classified by t_1/α methods.

5. *Finite strain estimates.*

During the past few years the emphasis in structural geology has shifted from the mere description of structures to an analysis of the stress, strain and mechanisms of deformation. One aspect that has received considerable attention is the determination of the amount of type of finite strain to which deformed rocks are subjected. This is most easily accomplished with objects of known original shape such as fossils (Breddin 1964; Tan 1973) but it has been attempted in most cases with conglomerate pebbles or oolites (Flinn 1956; Anhaueusser 1966; Hossack 1968; Dunnett 1969; Gay 1969; Elliott 1970; Mathews et al. 1974). It has also been attempted with crystals (Wilson 1972; Durney and Ramsay 1973; Mukhopadhyay 1973) and in one case (Borradaile 1974) with neptunian dykes.

The use of shear zones to determine finite strain states has been reviewed in the last chapter.

Modern finite strain analyses using folds have received relatively little attention (Ramsay 1969, p.58; Mukhopadhyay et al. 1969; Talbot 1970; Hudleston 1973c; Coward 1973a) possibly because of the many untestable assumptions inherent in the method (Hudleston op. cit. p.126). However, folds are far more widely developed in highly deformed rocks than are conglomerates and the writer believes that the possibility of making strain estimates is so rare that every attempt must be made to use the available material.

The method of investigation involves the recognition of 3 different processes operating during the deformation of layered rocks. These are the pre-fold layer-parallel shortening (Sherwin and Chapple 1968), buckling, and post-fold homogeneous flattening (Ramsay 1967, pp.411-415). The strain analysis requires that these deformations be 'removed' starting with the last.

All class 1C folds (the most common class in the Onseepkans area) may be regarded as flattened parallel folds and the axial ratio of the post-fold pure shear strain ellipse may be read directly off a t_1/α graph (Ramsay 1967, Fig. 7-79). The higher the flattening strain the more the fold approaches class 2. The folds may be unflattened by graphical means and the dihedral angle measured. The axial ratio of the strain ellipse required to produce this dihedral angle from a fold with limb dips of 15° may then be computed since it has been shown by Hudleston (1973b) that fold amplification from 15° limb dips takes place without change in arc length.

The initial stage of deformation producing folds with 15° limb dips must also take into account the layer-parallel shortening and this can be estimated using the graph in Fig. 73 taken from Sherwin and Chapple (1968). A review of these methods is presented in Hudleston (1973c, pp.121-126).

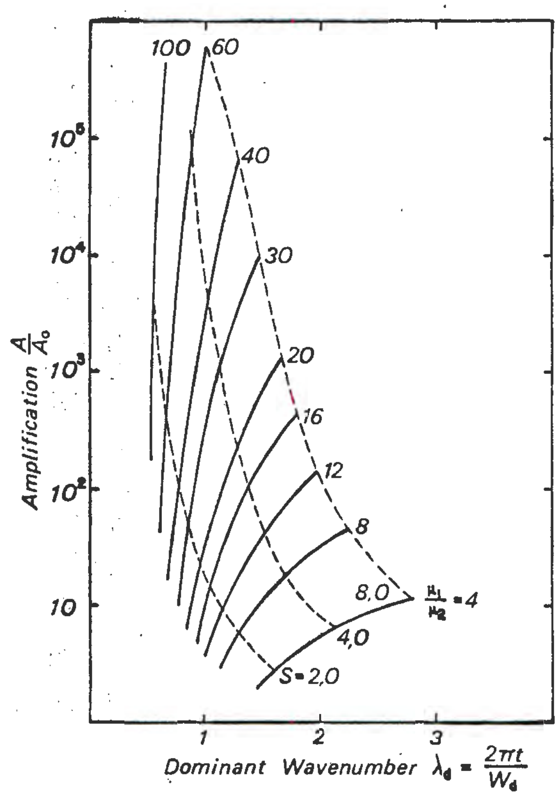


Figure 73. Graph for the estimation of strain resulting from layer-parallel shortening and buckling for folds with limb dips up to 15°. From Sherwin and Chapple (1968, Fig. 5).

6. *Approach to Fold Analysis in the Onseepkans Area*

It is not possible to give a comprehensive fold description of the 5000 km area covered in this report. The writer consequently decided to limit the study by examining folds from specially selected areas.

It has been pointed out by Elliot (1972) that folds growing in a plane strain environment (fold axis parallel to Y) have hinge zones that are the site of coaxially-accumulating deformation paths, whereas in the fold limbs new material surfaces become the principal strain sites with each increment of strain. This statement is easily verifiable in many cases since hinge areas of major folds are the locus of abundant minor parasitic folds whereas parasitic folds on the limbs of major structures are comparatively few.

In the field minor folds show a pronounced increase in numbers in the hinge zones of large structures and therefore minor folds in these areas may be taken as typical of the deformation pattern that produced the larger structures. The writer selected for analysis mesoscopic folds taken from the hinge zones of several major folds found in the eastern and western sub-areas. In practice the fold analysis is confined to D_3 and D_4 folds since minor folds associated with other deformational events are either absent (D_1, D_2) or very rare (D_5, D_6). In the following sections the minor folds are described by generation in the relevant sub-areas.

B. MESOSCOPIC FOLDS IN THE WESTERN SUB-AREA

1. *D_3 folds, classification and shape factor studies.*

The largest D_3 fold is the Leopard synform and the majority of minor folds described in the western sub-area come from its hinge zone. Fig. 74 illustrates the variety of the fold types found in this locality (c.f. also Plate 19 and Fig. 40).

The most common rock type in the hinge zone is migmatised biotite gneiss and the majority of the analysed folds are defined by gneissic layering or migmatite neosomes. This adversely affects the interpretation of the data since metamorphic layering is seldom mineralogically consistent and the composition of an individual horizon may vary considerably from one end of the fold strain to the other. There is a strong possibility that many of these horizons were not planar prior to folding and this may also lead to erroneous interpretations. In some cases (Plate 47) the anastomosing nature of the neosomes is obvious but in most cases the superimposed flattening strain has obliterated initial irregularities in the folded layers.

About 80 quarter-wavelength folds were studied by the dip isogon method. In most cases it was found that the isogons diverged from the inner to the outer arc of folds defined by quartzo-feldspathic horizons and converged on the outer arc of folds defined by biotite-rich layers (Fig. 75).

In some instances isogons are very nearly parallel to the fold axial planes but even when this occurs in one part of a fold train the 'normal' converging/divergent pattern is detectable in the other parts. On t_1/α plots it was found that the quartzo-feldspathic horizons tended to give class 1C folds and the mafic-rich horizons class 3 folds. Fig. 76 is a t_1/α plot of the folds in Fig. 75 and illustrates a well developed example of this phenomenon. The overall effect in some fold trains was a class 2 form produced by alternating classes 1C and 3 folds.

These plots and the dip isogon patterns show that the quartzo-feldspathic layers have behaved more competently than the biotite schists, and similar conclusions have been reached in other studies of this nature by Hudleston (1973c, p.117) and Coward (1973a, p.144). Thus, although the total effect may be a class 2 form, i.e. similar folds (Ramsay 1967, pp.430-436), the analysis has shown that the structures have not developed regardless of the lithology to produce what Wynne-Edwards (1963) has called flow folds or what Donath and Parker (1964, Plate 6) called passive folds (Hudleston 1973, p.112).

Dip isogon studies and t_1/α plots were also made on folds in amphibolite horizons but results were less consistent than above. Where amphibolites are bounded by biotite schist the folds show dip isogons convergent on the inner arc of the fold and the t_1/α plots show the amphibolite to be more competent (class 1C) than the biotite schist (Fig. 77 and Plate 48).

Folds defined by amphibolites in quartzo-feldspathic rocks, however, show dip isogons both convergent and divergent on the inner arc of the fold (Fig. 78A). A t_1/α plot of one horizon shows a clearly defined class 1C fold (Fig. 78B). This anomolous behaviour of the hornblende-bearing rocks may help to explain the variability of the fabrics seen in amphibolites from the eastern sub-area (c.f. Fig. 29).

In the Leopard synform hinge zone the sequence of rock types in terms of increasing ductility are: quartzo-feldspathic gneiss $<>$ amphibolite $>$ biotite schist. Both Coward (1973, p.144) and Francis (1973, p.177) found that amphibolites are less competent than quartzo-feldspathic rocks and therefore the present ductility sequence is different in this respect.

It can be seen from Fig. 74 that the shape of D_3 minor folds covers a wide range and this variation may be found in single fold train. In Plate 49 for example the first order antiform is a relatively open structure with a rounded fold hinge but the associated synform (below the scale) is isoclinal and has a sharp hinge area. To the above right of the scale small second order folds are isoclinal with sharp hinges and are disrupted in some cases to form minute rootless or intrafolial folds.

For purposes of comparison the shape factor of folds may be qualified using harmonic analysis and for this purpose 63 quarter-wavelength folds were analysed. Since the majority of the folds in this locality are defined by migmatite banding or metamorphic layering it was not possible to gather sufficient data for shape analysis by different rock types.

The magnitude of the first harmonic (b_1) indicates the wavelength/amplitude ratio or the degree of fold tightness. A frequency histogram of for the Leopard synform minor folds (Fig. 79) shows an asymmetrical distribution about a mean of 3,2. This figure may be compared with published 'shape factor' charts by Stabler (1968) and Hudleston (1973a) and in terms of the Fleuty (1964, p.470) classification the average minor fold would be described as 'tight'.

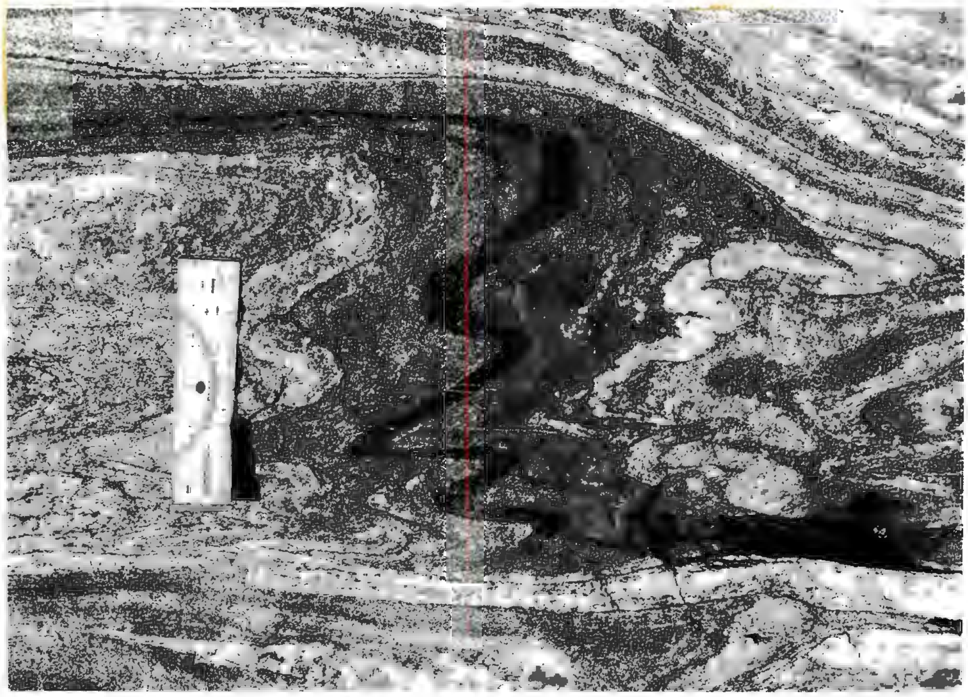


Plate 48. Amphibolite horizon in biotite schist forming class 1C folds. Keimasmond Farm.

Plate 49. A Leopard synform minor fold showing the variability of fold style within a single structure, Orangefall Farm.

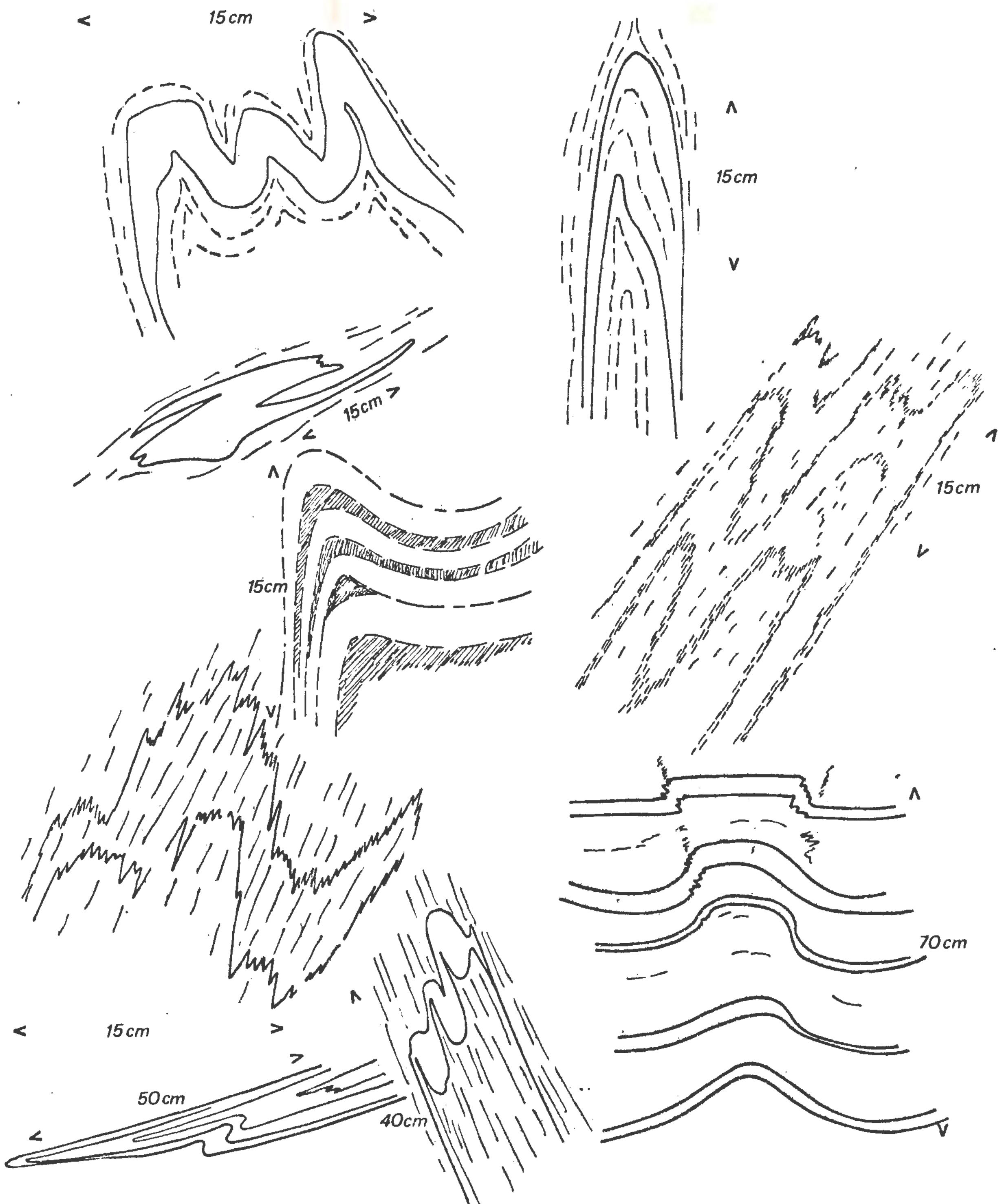


Figure 74. Drawings of mesoscopic folds taken from the hinge zone of the Leopard synform.

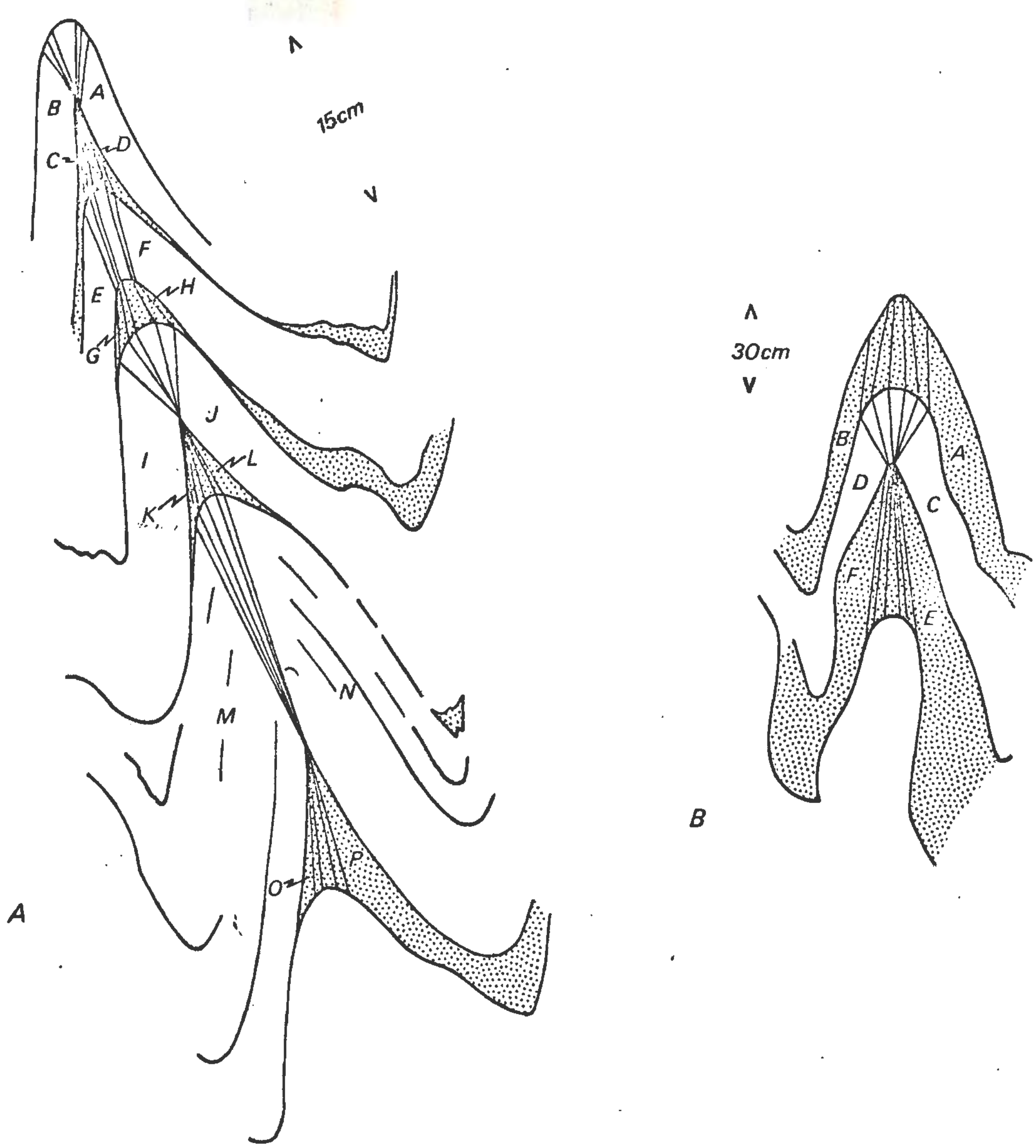


Figure 75. Dip isogons drawn on folds defined by alternating quartzo-feldspathic layers and biotite schist layers. Note the strongly convergent isogons in the outer arc of the biotite schist layers (stippled).

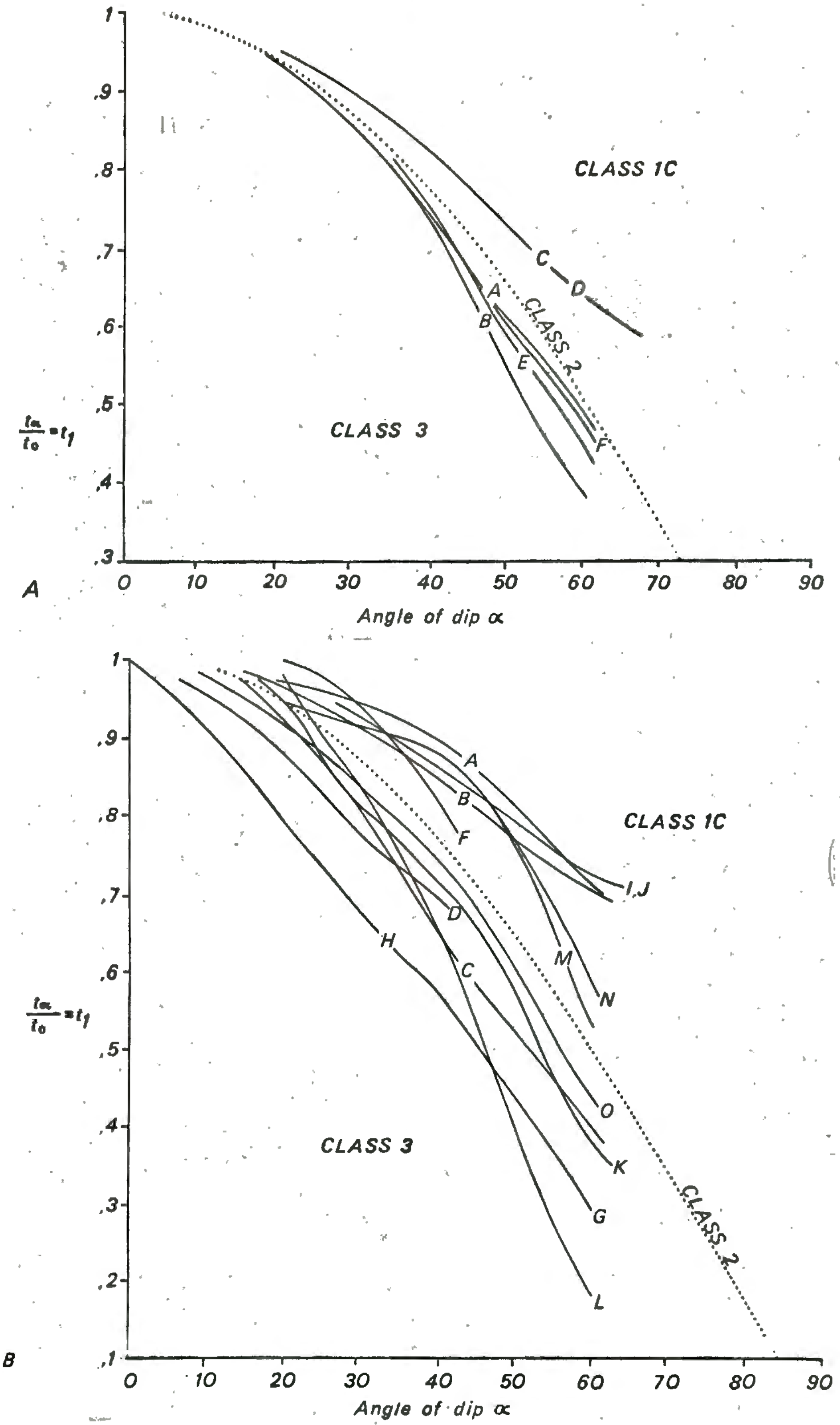


Figure 76. Classification of folds in Fig. 75. A and B refer to A and B in Fig. 75.

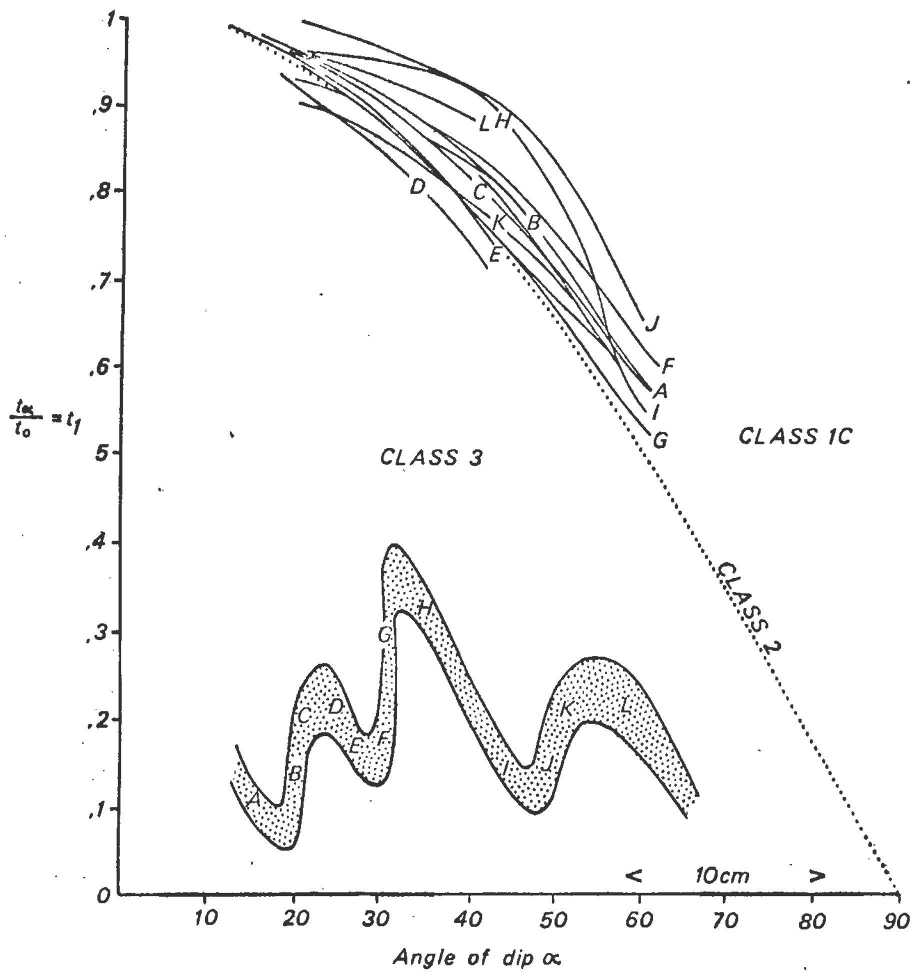


Figure 77. Fold classification of an amphibolite horizon in biotite schist (Plate 48).

The ratio of the third to the first harmonic (b_3/b_1) gives a quantitative estimate of the overall fold shape. Fig. 80A is a frequency histogram for b_3/b_1 and shows a symmetrical distribution about a mean of 0,06. The data were further broken down to give a comparison of the inner and outer arcs of folds having a class 1C geometry. The b_3/b_1 ratio for the inner arcs (Fig. 80B) gives a mean of 0.04 while the outer arcs have a mean of 0.08 (Fig. 80C). This implies that the inner arcs of an average class 1C fold have sharper hinges than the outer arcs. This phenomenon is well illustrated in Plate 50 and in Fig. 81A which is a b_1/b_3 graph for this pegmatite. The plots for the inner and outer arc describe different trends as a result of their 'sharpness' contrast between inner and outer arc. The data therefore suggest that the class 1C folds may have been formed by a process of tangential longitudinal strain in which compression in the inner arc of the fold caused a

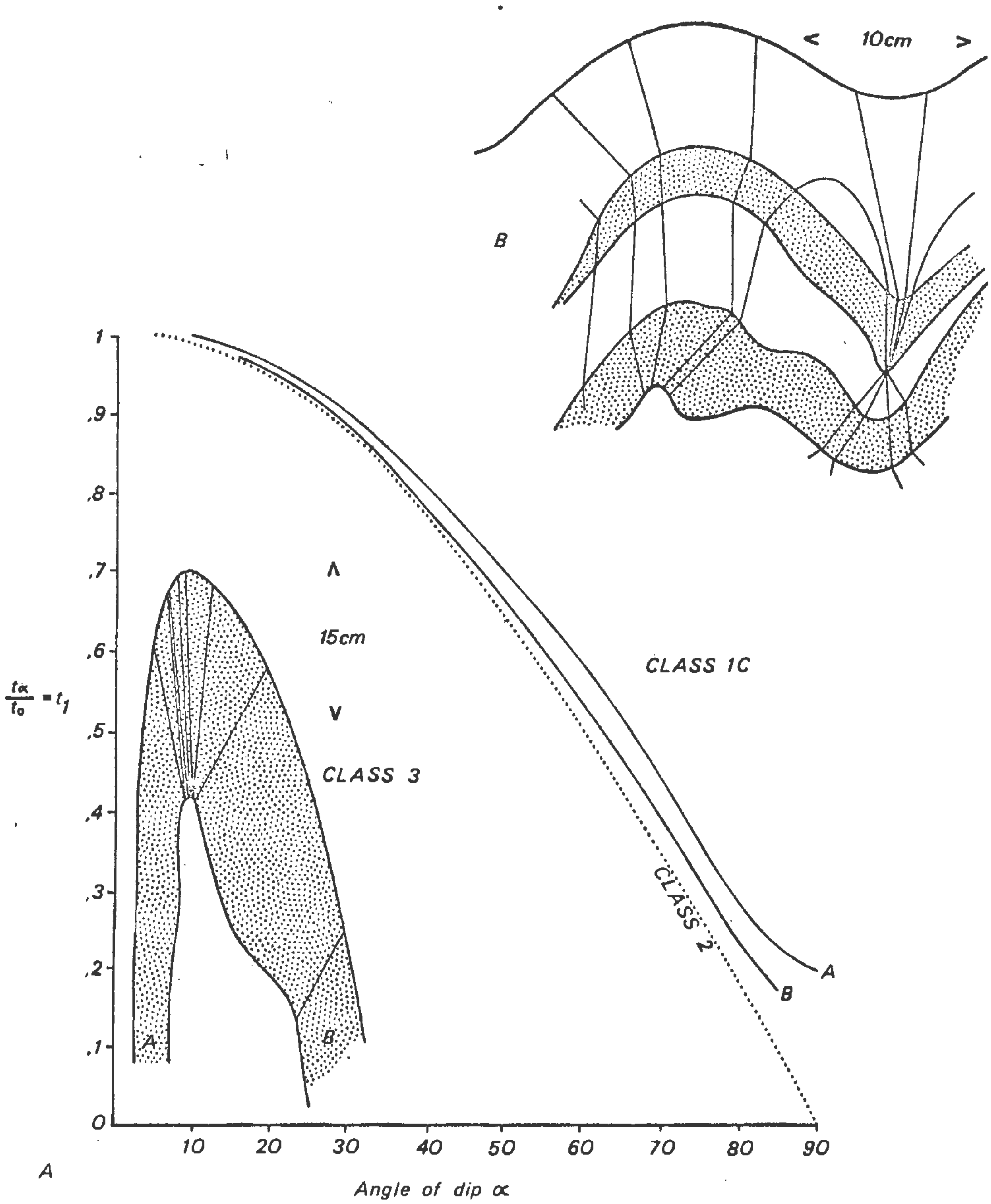


Figure 78. A. Dip isogons drawn on alternating amphibolite layers (stippled) and quartzo-feldspathic layers.
B. Classification of an amphibolite horizon in quartzo-feldspathic gneiss.

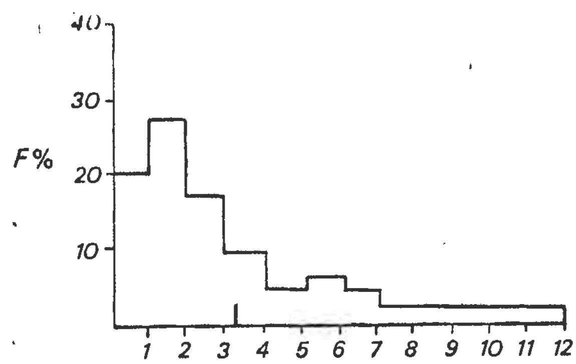


Figure 79. Frequency histogram for the first harmonic for 65 quarter wave-length folds in the Leopard synform. Distribution is asymmetric about a mean of 3,2.

sharp hinge while extension above the neutral surface produced a more rounded hinge zone (Fig. 81B).

A similar but less pronounced trend is shown by b_1/b_3 graphs of other folds. Most folds plot along the line $b_3 = 0$ with b_1 values up to 3. According to Hudleston (1973c, p.119) this distribution results from the flattening of gently buckled layers.

2. D_3 *fold finite strain estimates.*

Finite strain estimates were made on 20 folds (Fig. 82) according to the method outlined in section (A5) and the total shortening suffered by the rocks in the fold profile plane has a mean value of $\sqrt{\lambda_2/\lambda_1}=\cdot 11$ which represents a strain ellipse with axial ratios of 9:1. The bulk of this figure is given by the pre-buckle layer-parallel shortening as estimated from the chart (Fig. 73) given by Sherwin and Chapple (1968). The mean pre-buckle strain is $\sqrt{\lambda_2/\lambda_1}=0{,}33$ (ellipse 3:1) and the mean flattening strain $\sqrt{\lambda_2/\lambda_1}=0{,}54$ (ellipse 1,9:1) and the buckling strain is $\sqrt{\lambda_2/\lambda_1}=0{,}64$ (ellipse 1,6:1). These figures are very similar to those of a study made by Hudleston (1973c, p.124) on a much larger selection of minor folds in the Scottish Highlands.

Some problems were encountered with the pre-buckle strain computation as the distribution about the mean is much more asymmetric here than in either the flattening or buckling strains. This led to total finite strains recorded in some folds as high as 87:1 which appears to be an unrealistic figure. This

Plate 50. Folded pegmatite vein in the Leopard synform illustrating the shape difference between inner and outer arcs of class 1C folds. Orangefall Farm.

Plate 51. Bushman Hills antiform minor fold refolding D_3 structures (seen at both ends of the scale) and illustrating how earlier deformation may affect fold classification. Velloorsdrift Farm.

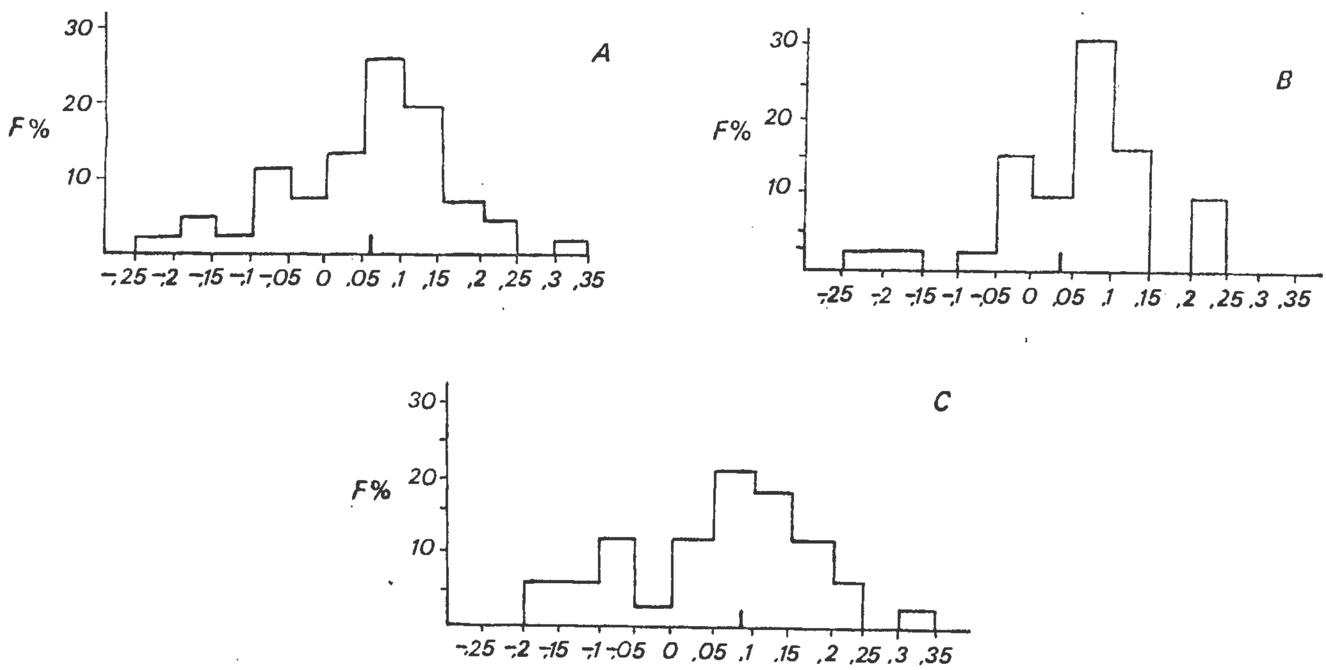


Figure 80. A. Frequency histogram for the ratios of the third to the first harmonic (b_3/b_1). Distribution is symmetric about a mean of 0,06. Total 65 quarter-wavelength folds.
B. Frequency histogram of b_3/b_1 for the inner arcs of class 1C folds, mean 0,04. Total 31 quarter-wavelength folds.
C. Frequency histogram for b_3/b_1 for outer arcs of class 1C folds, mean 0,08. Total 34 quarter-wavelength folds.

indicates that some of the assumptions used by Sherwin and Chapple (1968) may be open to question.

Some fold trains show little variation in flattening strain; $\sqrt{\lambda_2/\lambda_1}$ for the fold in Fig. 77 varies between 0,3 and 0,4. The flattening strains in the folded pegmatite (Plate 50 and Fig. 83), however, vary between 0,2 and 0,8 and illustrate strain heterogeneity over a very small area.

3. *D_4 folds; classification and shape factor studies.*

The best preserved D_4 fold in the western sub-area is the Bushman Hills antiform and minor folds from its hinge zone were collected from the area where its axial trace crosses the Velloor River valley (Fig. 84). Here the folds are again defined predominantly by gneissose foliation and migmatite

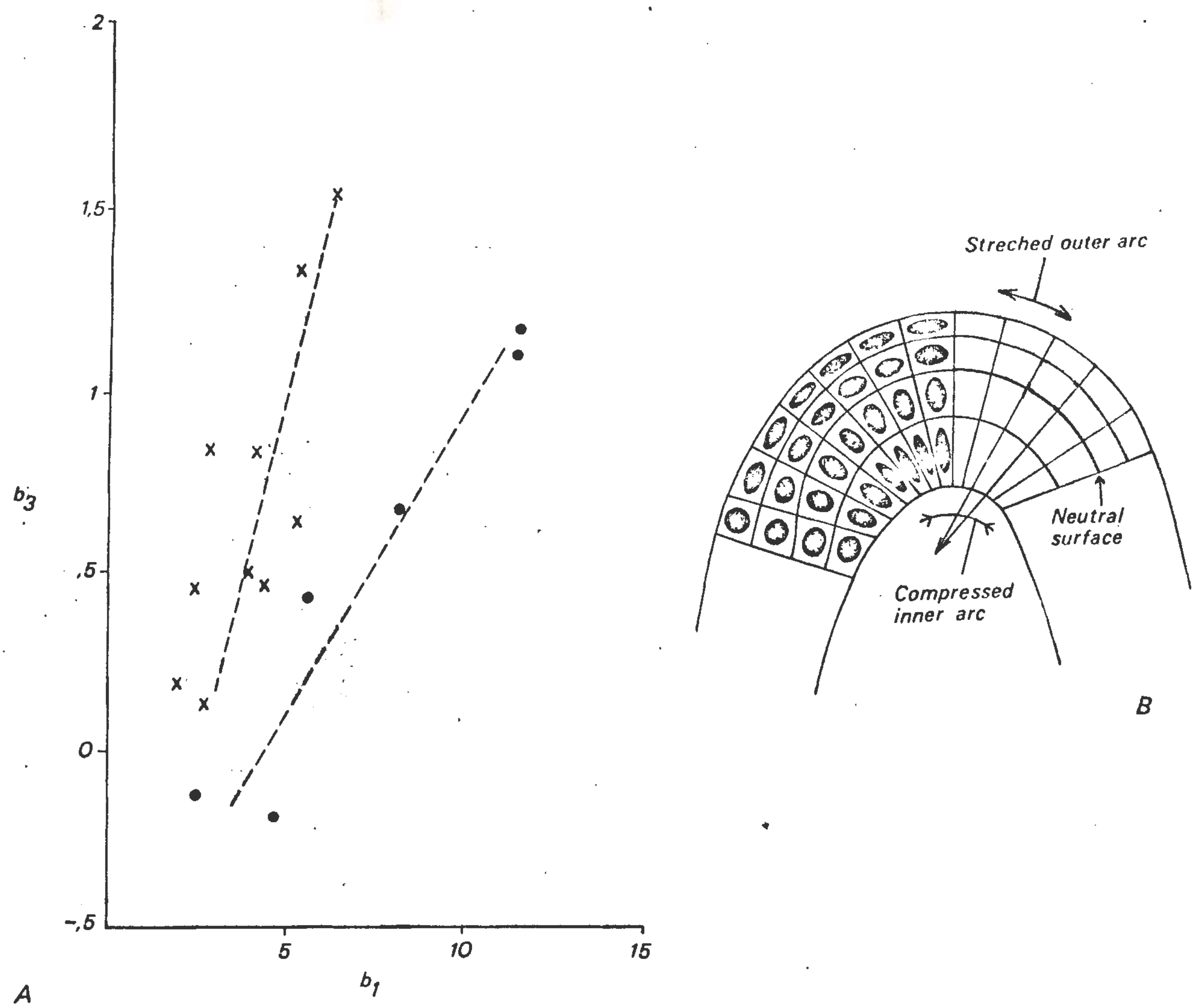


Figure 81. A. Graph of b_3/b_1 for the inner arcs (dots) and outer arcs (crosses) of the pegmatite shown in Plate 50. Dashed lines indicate linear trends of data.
B. Fold formed by tangential longitudinal strain which will give distribution of data as shown on the b_3/b_1 graph. After Ramsay (1967, Fig. 7-63).

neosomes, but amphibolites are more common in this locality than in the Leopard synform and a higher percentage are defined by alternating hornblende-rich layers and quartzo-feldspathic layers. Additional problems in interpretation are posed by refolded D_3 folds which can cause misleading variations in layer thickness and hence fold classification (Plate 51).

Dip isogon studies show that the patterns produced are exactly as that seen in the Leopard synform; the isogons converge towards the inner arc of

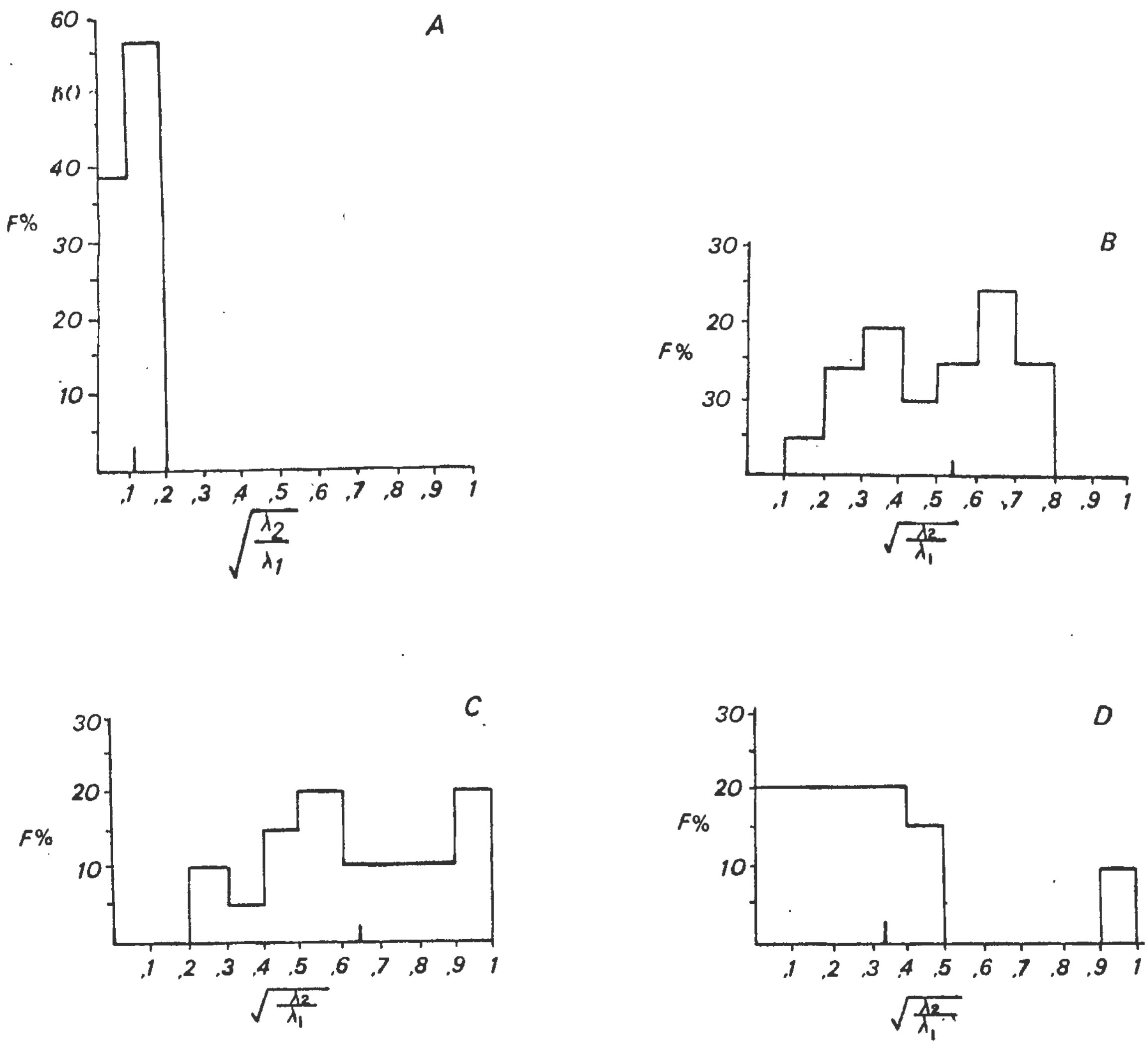


Figure 82. Frequency histograms for the various components of finite strain computed for 20 Leopard synform minor folds.
A. Total finite strain, mean 0,11
B. Flattening strain, mean 0,54
C. Buckling strain, mean 0,64
D. Pre-buckle layer parallel shortening, mean 0,33.

of quartzo-feldspathic layers and converge on the outer arc of micaceous horizons. On t_1/α plots the effect is one of alternating class 1C and class 3 folds although this may produce an overall class 2 appearance.

The anomalous behaviour of amphibolite horizons when compared to quartzo-feldspathic rocks is also apparent in D_4 folds. Dip isogons can either converge or diverge towards the inner arc of amphibolites, sometimes with both patterns in a single fold train and where amphibolite horizons are confined between quartzo-feldspathic rocks they often show either class 1C or class 3 folds (Fig. 85) but occasionally more complex patterns are found; in Fig. 86

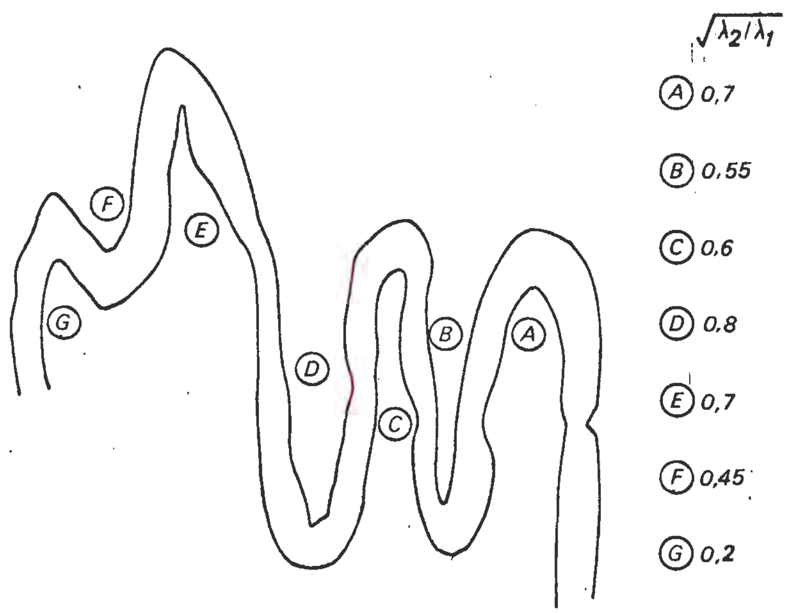


Figure 83. Variation in flattening strain through a single fold train (Plate 50).

amphibolite horizons (EF) on the inner arc of a quartzo-feldspathic layer (CD) give class 3 folds while the amphibolite horizon on the outer arc (AB) gives class 1C and 1A. According to Ramsay (1967, p.416) and Hudleston (1963c, p.117) a similar situation occurs in a zone of contact strain around buckled layers where class 3 folds are produced on the inner arc and class 1A folds on the outer arc. In the example shown in Fig. 85, however, the 'buckled' layer itself (CD) is class 3 - class 2 whereas class 1B or 1C would be expected. This may possibly be explained by a strong flattening strain having modified the fold geometry (Hudleston op. cit., p.118).

The frequency histogram for the first harmonic (Fig. 86A) has a mean of 2,9 and indicates that the measured D_4 folds (total 46) are very slightly more open than measured minor D_3 folds (mean 3,2). The mean b_3/b_1 is 0,06 and the distribution on the frequency histogram (Fig. 86B) is approximately symmetrical about the mean. The mean b_3/b_1 for the inner arcs of class 1C folds is 0,03 and for outer arcs 0,08 (Figs. 86C and 86D). These figures are virtually identical to those for D_3 folds and indicate both that the range of fold shapes and the possible fold mechanism is the same.

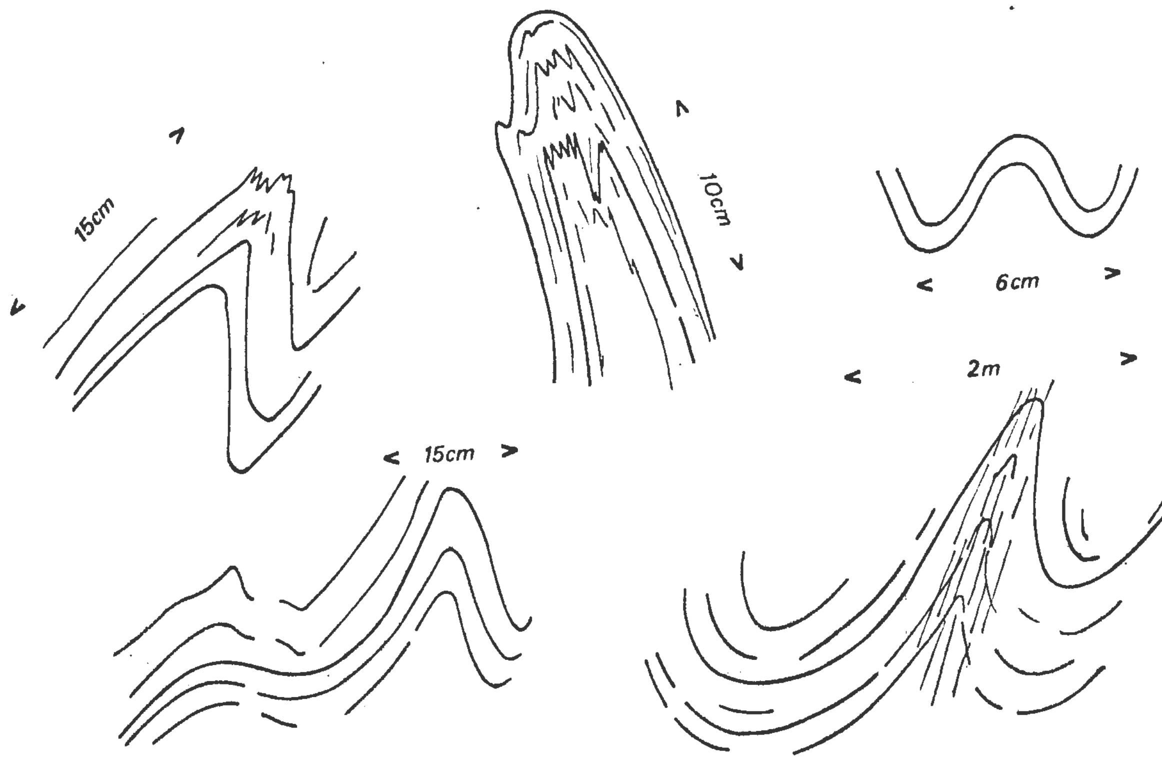


Figure 84. Mesoscopic folds from the hinge zone of the Bushman Hills antiform.

4. D_4 fold finite strain estimates.

The mean total finite strain in the fold profile plane measured on 17 folds is 0,16 (Fig. 87A) which gives an ellipse with axial ratio of 1:6,3. This compares with a mean of 0,11 (ellipse 1:9,1) for the measured Leopard synform folds. The breakdown of this finite strain figure into flattening strain, buckling strain and pre-buckling strain is shown in Fig. 87B, C and D. It can be seen that for all categories, the strains are lower than those for the D folds (compare with Fig. 82). The conclusion that D_4 folds have suffered less strain than D_3 folds is, to some extent, obvious since the earlier folds have had a much longer deformational history.

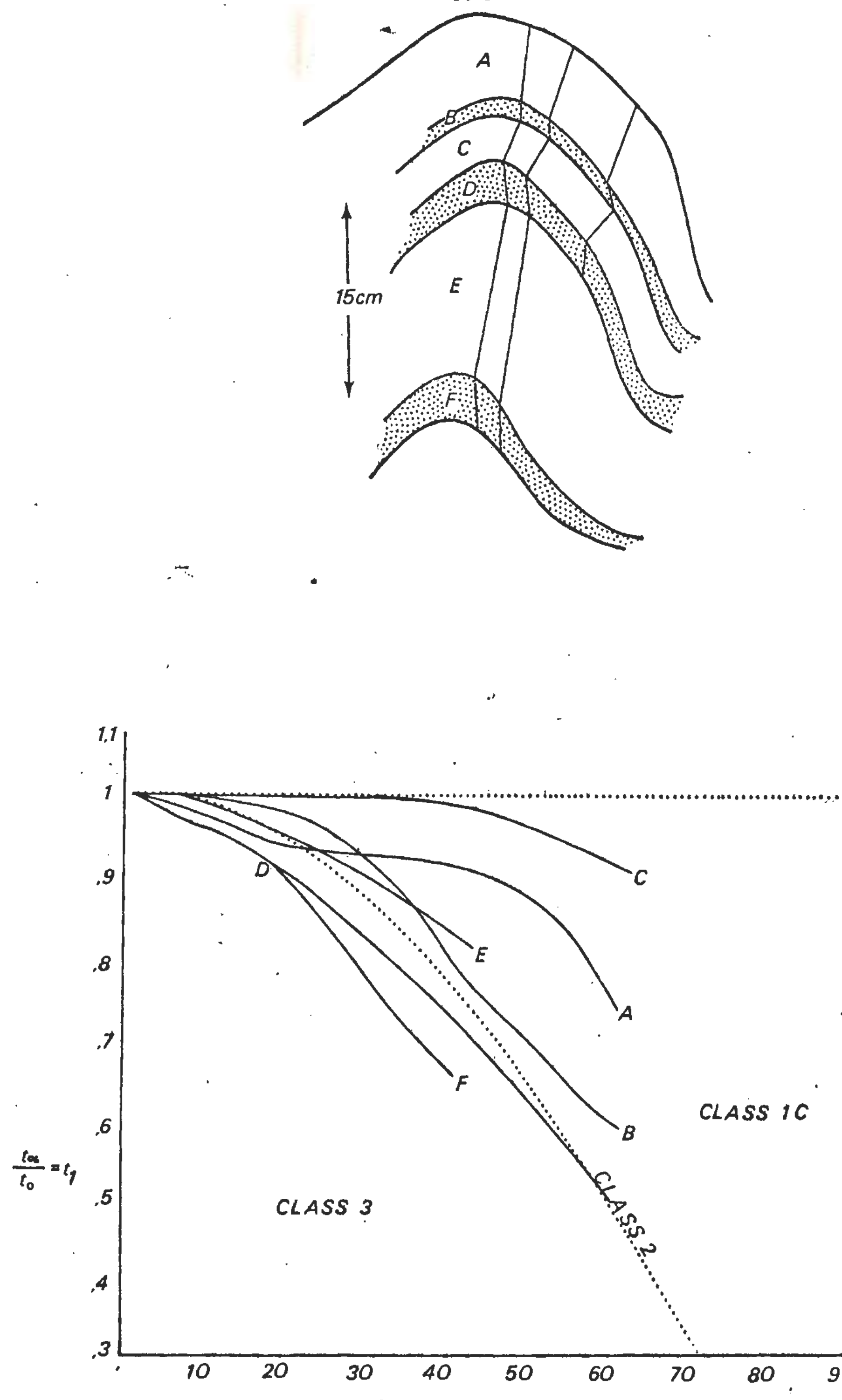


Figure 85. Fold classification of intercalated amphibolite and quartzo-feldspathic horizons in the Bushman Hills antiform.

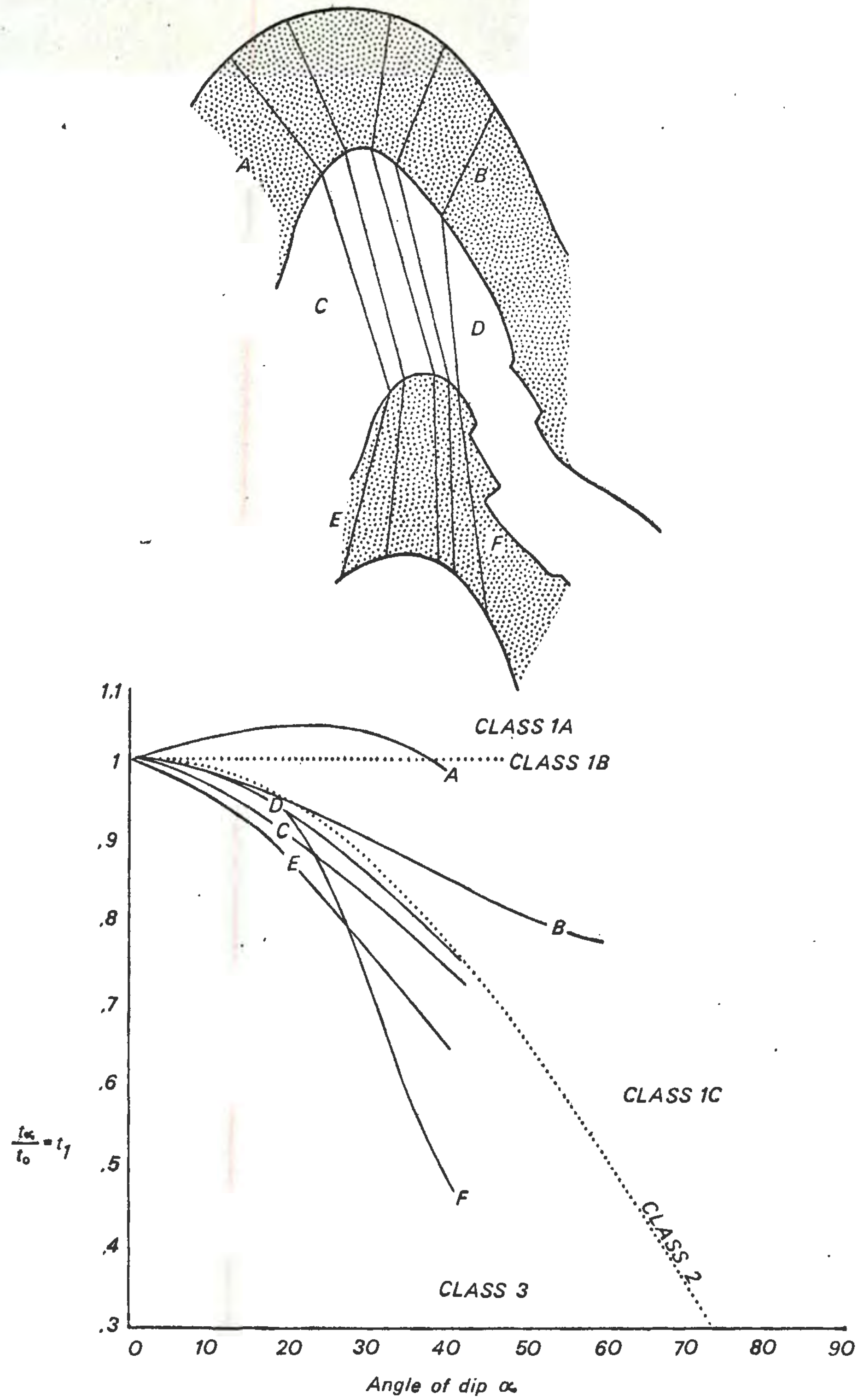


Figure 85. Fold classification for amphibolite horizons (stippled) on the inner and outer arc of a quartzo-feldspathic horizon (CD).

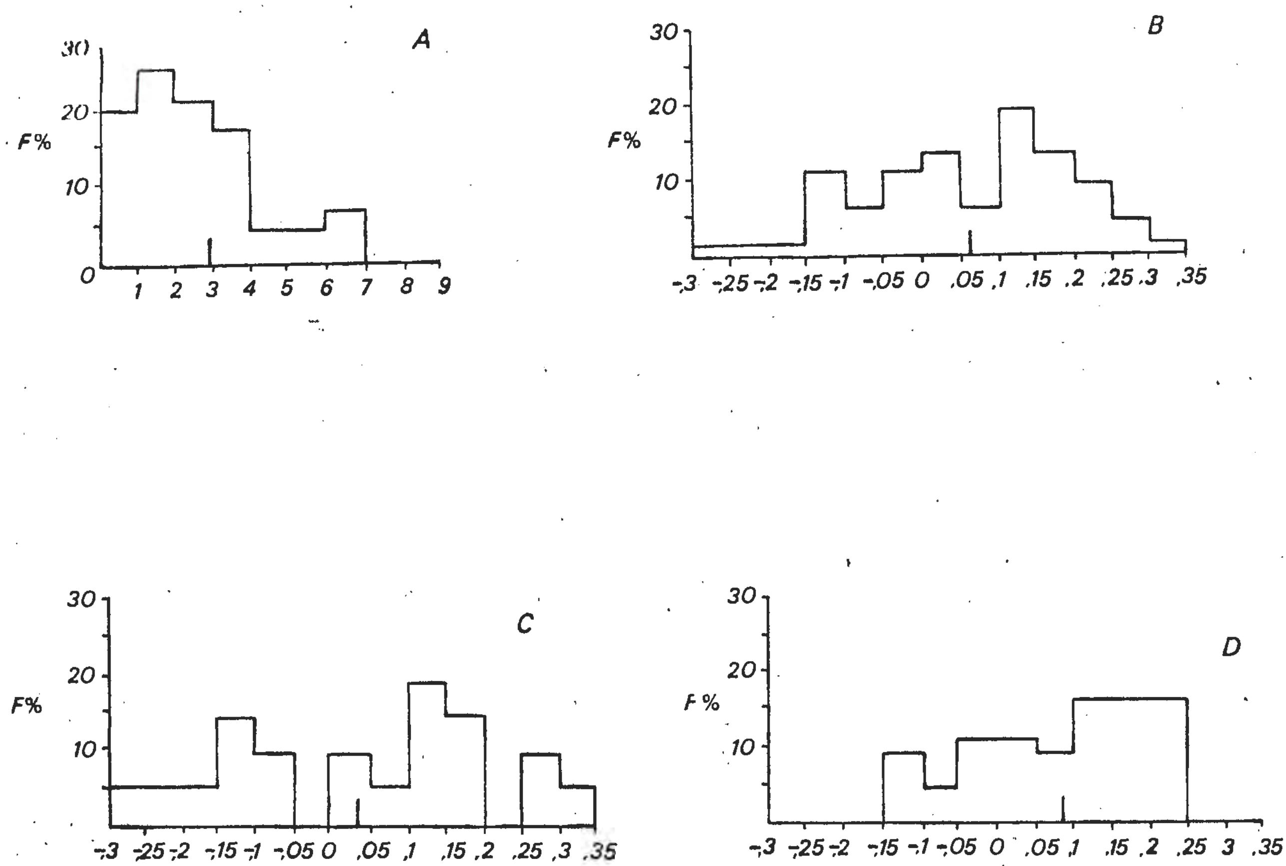


Figure 86. Harmonic analysis of mesoscopic folds in the Bushman Hills antiform.

A. Frequency histogram of b_1 for 47 quarter-wavelength folds, mean 2,9
B. Frequency histogram of 47 quarter-wavelength folds for ratio of b_3/b_1 ; mean 0,06
C. Frequency histogram of b_3/b_1 for 23 inner arcs of class 1C folds; mean 0,03
D. Frequency histogram of b_3/b_1 for 24 outer arcs of class 1C folds; mean 0,08

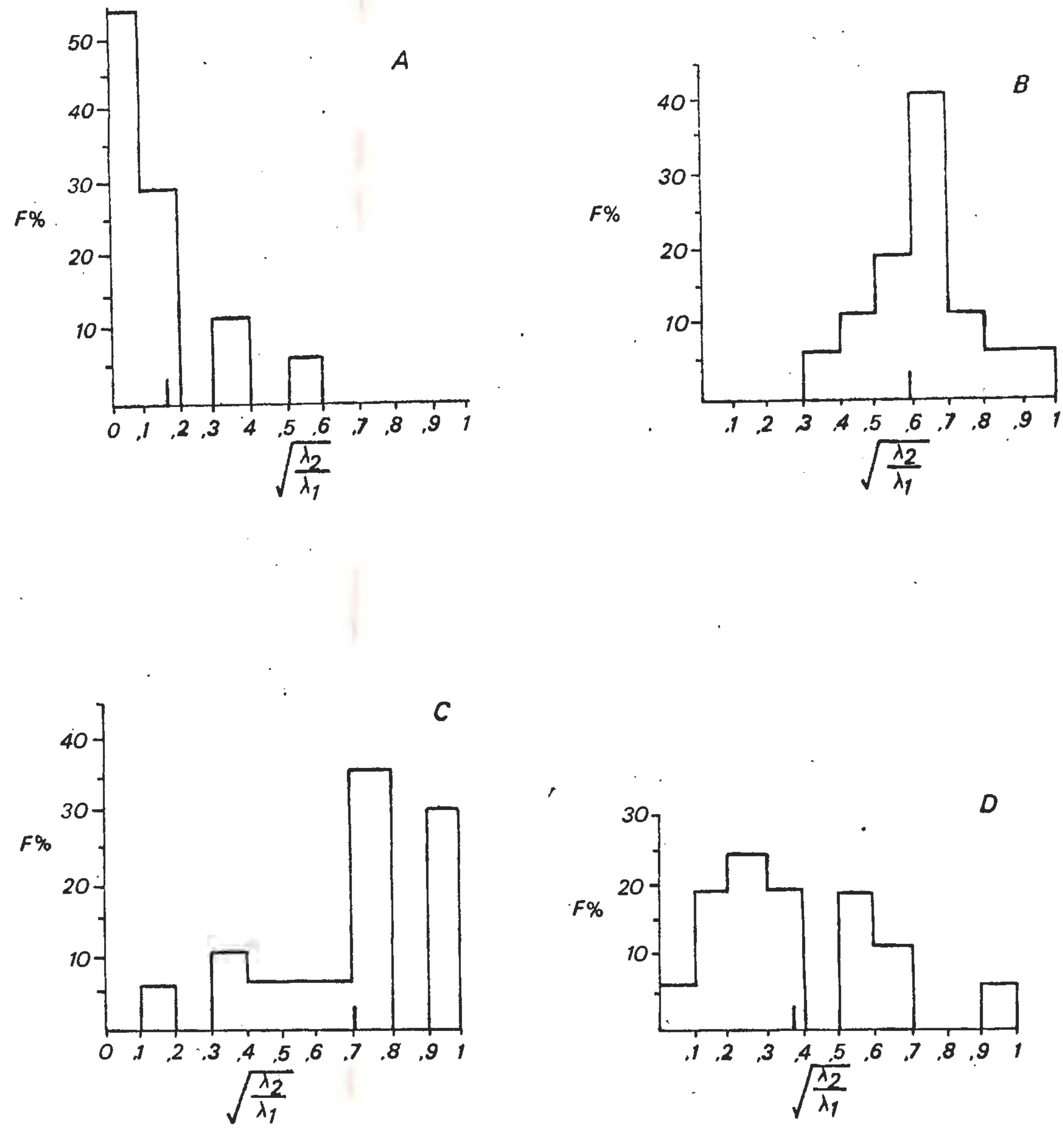


Figure 87. Frequency histograms of strain in 17 Bushman Hills antiform mesoscopic folds.

A. Total finite strain; mean 0,16

B. Flattening strain; mean 0,59

C. Buckling strain; mean 0,7

D. Pre-buckle layer parallel shortening; mean 0,37

C. THE EASTERN SUB-AREA

Data for mesoscopic fold description in this sub-area were collected from the hinge zones of the Plessis synform (D_3) and the Cobra synform (D_4) since it can be shown that these structures definitely belong to different generations.

1. *D_3 folds; classification and shape factor studies.*

The majority of the folds analysed in the Plessis synform are defined by quartzo-feldspathic layers and the mineralogical contrast between these layers is never great. This situation is very different from the D_3 Leopard Synform where the folds were mostly defined by alternating quartzo-feldspathic layers and biotite schist.

The dip isogon studies show that the convergent/divergent pattern between successive layers is still apparent but is not as pronounced as in rocks with contrasting mineralogy. The slightly more mafic-rich horizons show isogons weakly convergent on the outer arc of the fold and on t_1/α plots these horizons show a class 3 form very close to class 2. This confirms the differing rheological behaviour of rocks with higher mafic mineral content that was established in folds from the western sub-area. No photographs of folds defined by amphibolite horizons were taken.

The first harmonic has a mean of 1,6 (Fig. 88) and there is less variance about the mean than seen in the Leopard synform folds (Fig. 79). This would again appear to be the result of low compositional variations in successive layers in the fold. It has been suggested by Hudleston (1973c, Fig. 14) from a comparison of b_1 histograms for folds in different lithologies that the mean and the variance about the mean increases sharply as the mineralogical distinction between the layers increases.

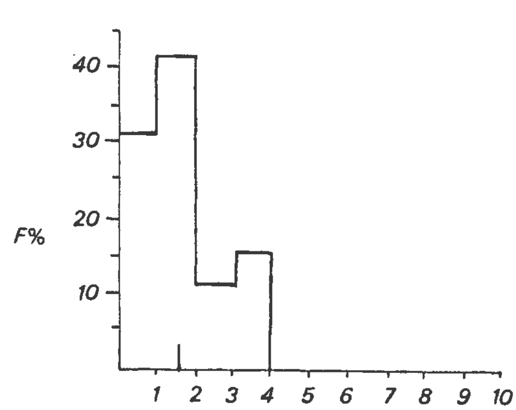


Figure 88. Frequency histogram for the first harmonic in 19 Plessis synform quarter-wavelength folds; mean 1,6.

The frequency histogram for the ratio of the third to the first harmonic (Fig. 89) shows a symmetrical distribution about a mean of 0,01. This is significantly lower than the 0,06 mean for Leopard synform folds and it has also been proposed by Hudleston (1973, Fig. 15) that b_3/b_1 means increase in proportion to the compositional distinction between the layers. The mean b_3/b_1 for inner arcs of class 1C folds is 0,02 and for outer arcs 0,04; the distinction of 0,02 for inner and outer arcs is only half that for the Leopard synform folds (0,04) and therefore the shape difference is less marked. This may imply that a higher component of tangential longitudinal strain was involved in the production of class 1C Leopard synform folds where obvious compositional differences are apparent. In the Plessis antiform the folds have perhaps developed by flexural slip on surfaces between quartzo-feldspathic layers of approximately the same competency and therefore the 'sharpness' of the inner arcs *vis* the outer arcs is not as noticeable.

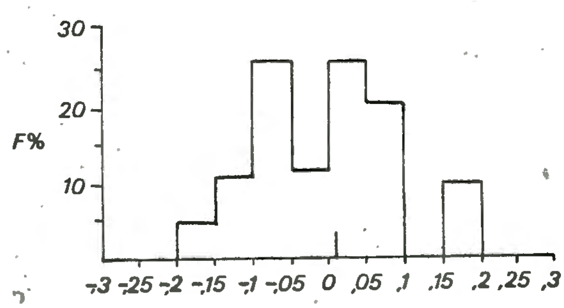


Figure 89. Frequency histogram for the ratio of the third to the first harmonic in 19 Plessis synform quarter wavelength folds; mean 0,01.

In many cases the folding produced in these quartzo-feldspathic rocks is related to the thickness of individual layers. Where, for example, a thin layer is sandwiched between two thick layers in a fold hinge, compound fold wavelengths are produced (Plate 52) and the thinner layers form second order folds (Ramsay 1967, pp.415-421).

2. *D_3 folds; finite strain estimates.*

Strain estimates were made on ten folds only and this may not be sufficient to make a meaningful comparison with D folds in the western sub-area. The total finite strain suffered by Plessis synform mesoscopic folds is $\sqrt{\lambda_2/\lambda_1}=0,11$ (ellipse 9:1) which is identical to the Leopard synform folds. The frequency histograms for the different strain components is shown in Fig. 90 and these are broadly comparable to the histograms in Fig. 82.

3. *D_4 folds; classification and shape factor studies.*

The rock type most commonly encountered at the southern end of the Cobra synform is deformed Beenbreek granite with remnants of pre-tectonic gneisses. Many of the folds analysed in this locality are defined by either aplitic material or migmatite neosomes. No folds defined by amphibolites were encountered.

Dip isogon patterns obtained on these folds show a strong convergence towards the inner arc of aplitic or neosome bands and on t_1/α plots these bands are classified as 1C folds with the surrounding augen gneisses forming class 3 folds. This relationship is very similar to the pattern produced by alternating neosome material and biotite schist in the Leopard synform.

The frequency histogram for the first harmonic (Fig. 91A) shows a higher mean than that for the adjacent Plessis synform and also a greater variance (compare with Fig. 88). The frequency histogram for the ratio of the first to the third harmonic (Fig. 88B) shows a symmetrical distinction about a mean of 0 which is, in fact, lower than would be expected from studies on the Leopard synform folds where the distinction between the layers is equally marked. However, the difference between the inner and outer arc is very marked; for inner arcs the mean b_3/b_1 is 0,06 and for outer arcs 0,03. This implies that the shape difference on the inner and outer arcs of class 1C folds is greater than any other major structure examined. It may be suggested that the relatively pure quartzo-feldspathic gneisses forming the aplitic material is therefore the most competent (least ductile) rock type in the area. Hudleston (1973, p.117) also found that pegmatites virtually free of mafic minerals provided the greatest contrast between inner and outer arcs of folds and he offered a similar interpretation.

Biotite schist horizons deformed between neosomes or aplitic material appear to have formed with a very high component of flexural flow. The limbs of these folds show a better foliation than the hinge zones and occasionally small neosome bands in these localities are strongly disrupted and form intrafolial folds on which the foliation is axial planar (Plate 53). The hinge zones, however, do not show strong second order folding and the axial plane foliation does not appear to be strongly developed (see also Fig. 43b).

4. *D_4 folds; finite strain estimates.*

Strain estimates from 11 folds gave a total shortening in the fold profile plane of $\sqrt{\lambda 2/\lambda 1} = 0,15$ (ellipse 7:1) as compared to $\sqrt{\lambda 2/\lambda 1} = 0,16$ for the D_4 Bushman Hills antiform in the western sub-area. The various components of finite strain are shown in Fig. 92 and these are very close to those obtained for the Bushman Hills antiform minor folds.

Plate 52. Plessis synform minor fold showing fold development with a high component of flexural slip in quartzo-feldspathic rocks. Compound fold wavelengths are developed in the fold hinge as a result of variable layers thickness. Stolzenfels Farm.

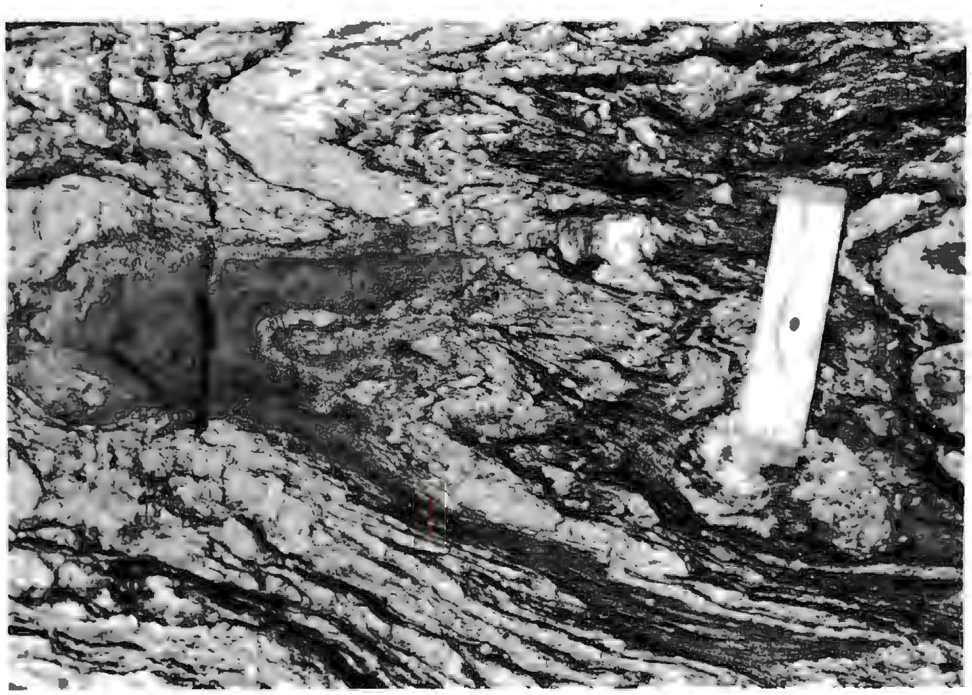


Plate 53. Flexural flow fold in biotite schist from the Cobra synform. Note well developed foliation and intrafolial folds on lower limb. Stolzenfels Farm.

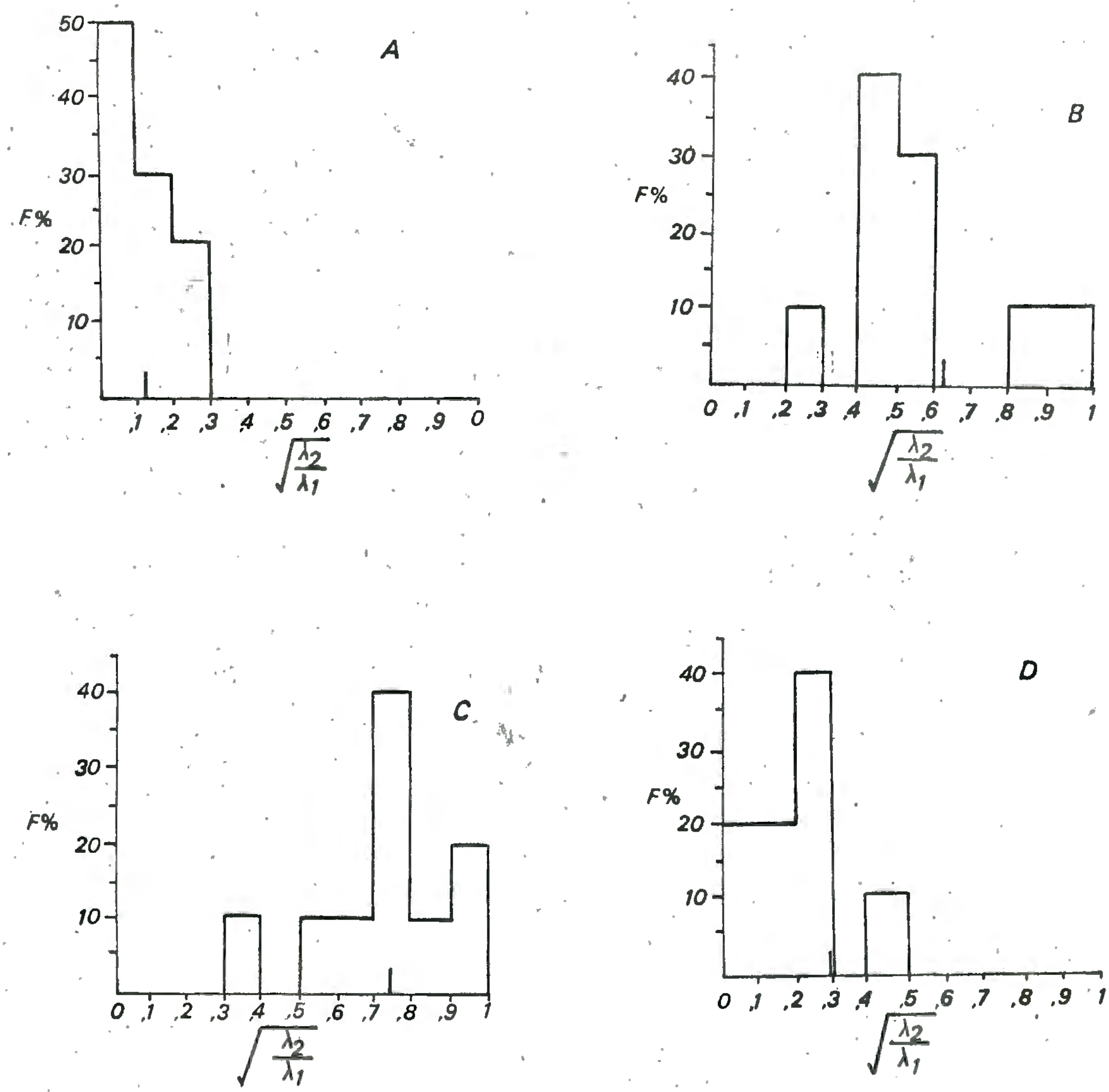


Figure 90. Frequency histograms for the various components of finite strain computed for 10 Plessis synform minor folds.

A. Total finite strain; mean 0,11
B. Flattening strain; mean 0,61
C. Buckling strain; mean 0,74
D. Pre-buckling layer-parallel shortening; mean 0,29

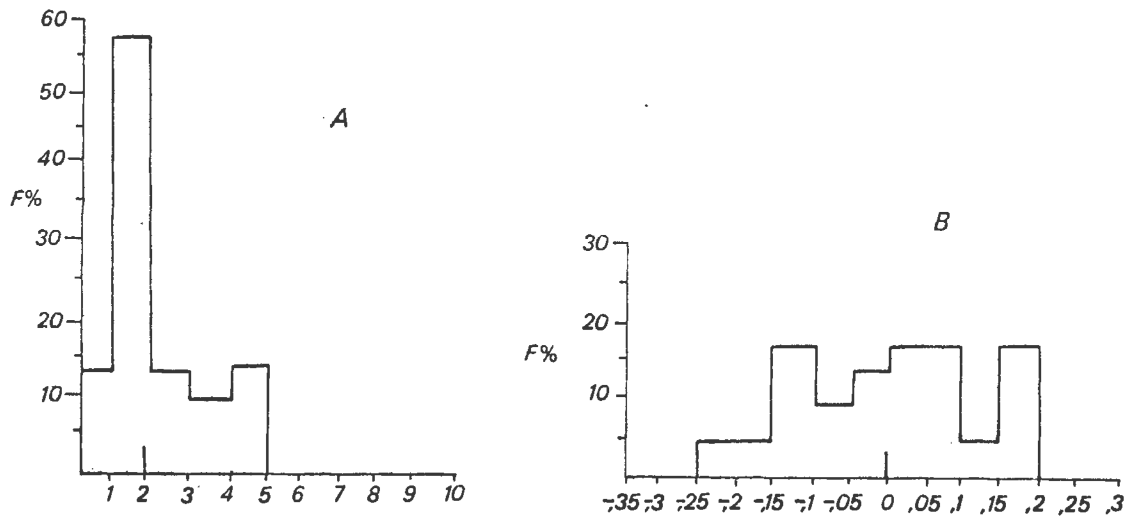


Figure 91. A. Frequency histogram for the first harmonic in 24 Cobra synform quarter-wavelength folds.
B. Frequency histogram for the ratio of the third to the first harmonic in 24 quarter-wavelength folds.

D. DISCUSSION

A selection of mesoscopic folds have been examined from the hinge zones of some of the major structures in the eastern and western sub-areas. It is assumed that minor folds in these localities are unambiguously related to the major structures and it is further assumed that these folds can be considered typical of the deformation producing the folds.

The last assumption is open to question as flexural flow folds will show higher strains of fold limbs. The difficulty in making positive correlation between major and minor structures in these localities, however, restricted the scope of the study.

Dip isogon studies and t_1/α fold classification have established certain patterns which suggest that the lithology defining the folds exercises a controlling influence over the class of fold produced. The pattern produced by an association of rock types is consistent not only over the entire area but also during succeeding deformation episodes. The only exception to this rule is the relation between amphibolites and quartzo-feldspathic gneisses where the former can either form class 1C or class 3 folds. However, this exception is also consistent in that the same variability continues through the second episode of deformation in the western sub-area.

The normal pattern that has emerged is that quartzo-feldspathic rocks form class 1C folds while intercalated more biotite-rich horizons form class 3

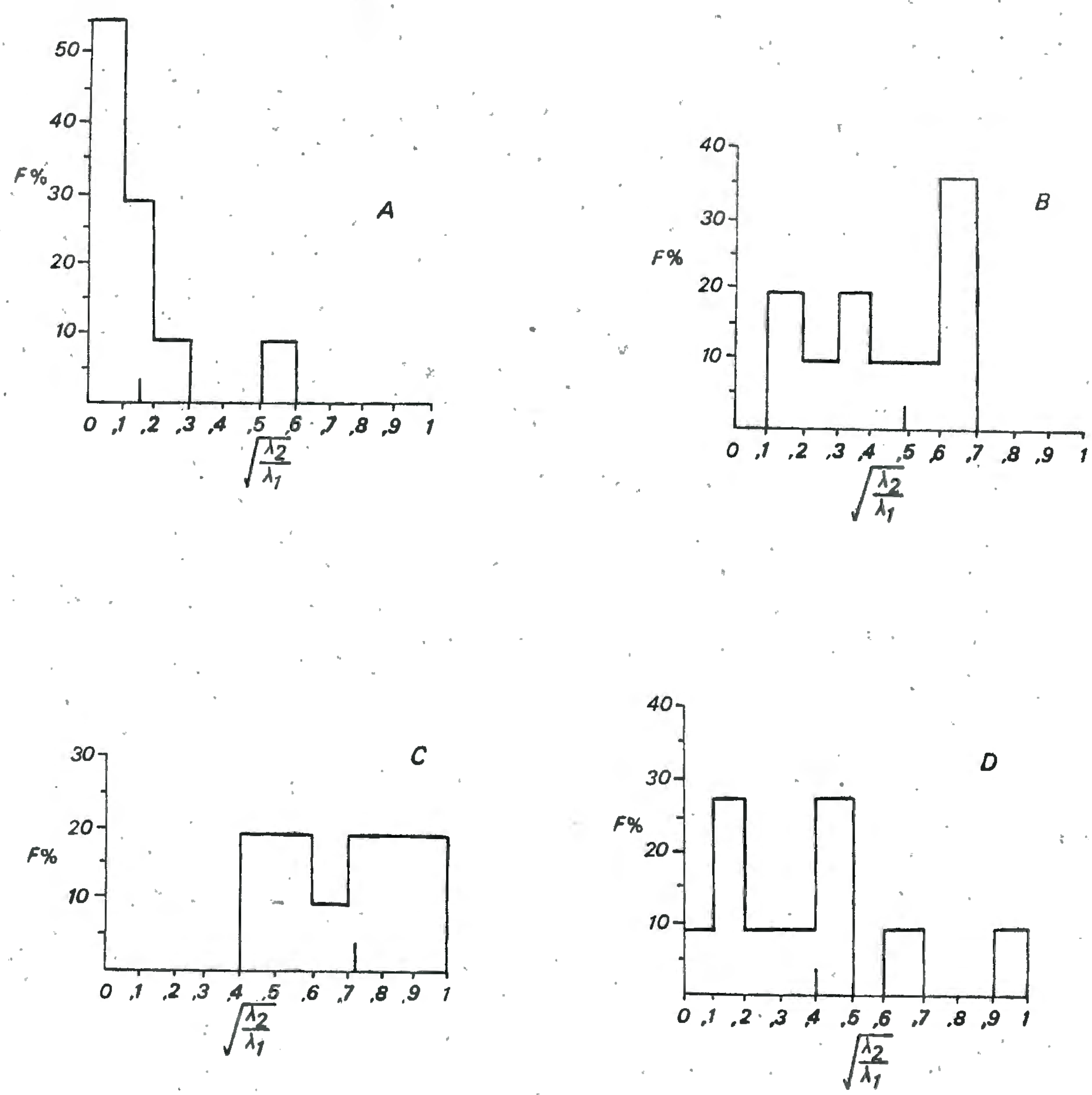


Figure 92. Frequency histograms for various components of finite strain measured on 11 Cobra synform minor folds.

A. Total finite strain; mean 0,15

B. Flattening strain; mean 0,49

C. Buckling strain; mean 0,71

D. Pre-buckle layer-parallel shortening; mean 0,4 folds. Quartzo-feldspathic rocks are therefore interpreted as being more competent than the biotite-bearing rocks.

Harmonic analysis has shown that definite shape differences exist between inner and outer arcs of folds. In many cases this indicates that class 1C folds (the majority of measured folds) developed by buckling and that the mechanism of deformation included a high component of tangential longitudinal strain. The layers have not behaved passively therefore and similar or 'shear' folds are not encountered.

One further aspect that may be emphasized is that the shape and tightness of measured folds shows no meaningful differences between folds of succeeding generations. In fact, the data presented here are virtually identical to a much more extensive study of minor folds made by Hudleston (1973c) in the Scottish Highlands.

A strain analysis has shown that a variable degree of homogeneous flattening has been superimposed on the buckle folds. Total finite strains in the fold profile plane between different fold generations vary between 6:1 and 9:1 and are broadly consistent for the whole area. These results are similar to other studies made on 'early' folds in other areas. Hudleston (1973c, p.124) found a total shortening of 11:1 for 157 F_2 folds in the Loch Monar area of the Scottish Highlands and Coward (1973a, p.146) has shown that 43 F_3 folds in the Laxfordian gneisses of South Uist have total strains varying between 7:1 and 14:1. The variance about the mean is high for most components of strain and some folds record strains of over 30:1. The histograms for flattening strain presented by Coward (op. cit. Fig. 4) show a similar variance.

One of the aims of structural geology is to provide a quantified account of deformation (finite strain) in naturally deformed rocks. The analyses discussed above give an indication of the finite strain represented by folds of different generations in different parts of the area. Such studies could be extended to other structures commonly found in the Namaqua belt such as deformed megacrysts and mafic inclusions in granite, deformed conglomerates and deformed porphyroblasts. Eventually a quantitative estimate of the finite strains represented by granites, shear zones, folded rocks etc. would lead to an estimate of the total shortening across the mobile belt. This aspect of structural geology, therefore, is one with far-reaching implications.

E. CORRELATION WITH OTHER PARTS OF THE NAMAQUA BELT

The polyphase nature of the deformation in the Namaqua belt is now well documented (Joubert 1971, 1974a, 1974b; Vajner 1974; Blignault et al. 1974; Jackson 1976; Beukes 1973) and the possibility of correlating deformational events between these areas is suggested. Such an approach has been strongly advocated by Hopgood (1973) and Hopgood and Bowes (1972) but the methods which these authors used have been criticised (c.f. Ramsay, Coward, Sutton and Windley in discussion of Hopgood 1971, and Coward in discussion of Bowes <u>in</u>

Francis 1973, pp.183-185). It would now appear that such correlations are invalid unless there is at least a well established continuity of major structures between the areas under comparison.

The recognition of a particular fold style by which individual mesoscopic folds of one generation can be recognised has been criticised by Park (1969), Williams (1969) and Ramsay (1967, p.546). The quantitative studies of folds belonging to different generations described above have furthermore shown that there is a considerable overlap in many features generally referred to as fold style.

The importance of layer thickness and the lithology in the development of the fold-forming mechanism has been emphasized in preceding sections. For example, thin component layers in a fold developed by flexural flow will form tight to isoclinal parasitic folds and sometimes intrafolial folds on the limbs (Plate 53) while in the hinge zones of the large structure open folds are present. Coward (1973a) has also shown that the orientation of axial planes of parasitic folds around a large structure may vary by up to 180^{o} since the direction of maximum shortening in a layer folded by tangential longitudinal strain has a similar variability (c.f. Fig. 81B).

The correlation of intrafolial folds might be particularly misleading. It was shown in Fig. 45 that these structures can result from normal progressive deformation in certain parts of a fold train; however, their ubiquity in the Onseepkans area requires an alternative mechanism.

It has been shown by Ramberg (1955) that boudinage results from compression perpendicular to the layering and the scale of boudin development is a function of the thickness of the competent layers and their viscosity contrast with the matrix. If a series of folds is flattened then the hinge zones of minor parasitic folds on the flanks of the larger structures have effectively 3 times the thickness compared to the remainder of the folded layer (Fig. 93). Boudinage will then result in these hinge zones becoming isolated to form intrafolial folds. If the layers are of varying thicknesses, it is quite possible for intrafolial folds to occur between thicker, apparently 'undeformed' layers (Fig. 93B and Plate 54). In the Onseepkans area folds of *two* generations have suffered a post-fold homogeneous flattening strain and, therefore, this mechanism of intrafolial fold formation was clearly available. The much quoted distinction of Rast (1956) between boudinage and rootless folds is therefore misleading; many rootless folds *are* boudins.

In most deformed rocks it is very difficult to determine the deformation path by which the finite strain state was obtained (Ramsay 1969, p.54) but the influence of the deformation path in terms of the structures developed in the rock has very important implications (Elliot 1972, p.2629). In the Onseepkans area a clear indication of this importance is given by a consideration of the Pofadder ZAHNCAFS, a feature known to be produced by predominantly rotational (non-coaxially accumulating) strain. A series of large folds produced in the ZAHNCAFS has a 90^{o} variation in orientation between the core and the margin. Adjacent to the core the folds are isoclinal and, in one locality, they have an axial plane foliation. At the margin of the zone some 30 km away these folds are open dome-and-basin structures with no accompanying fabric element. It can be seen from this example that a high component of heterogeneous rota-

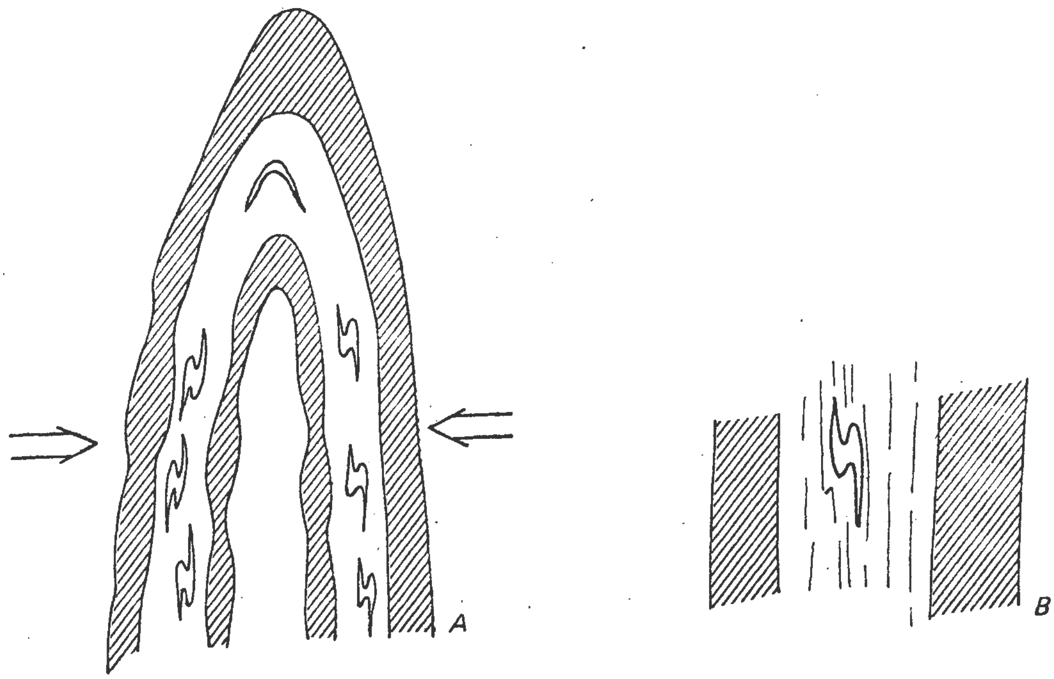


Figure 93. A. Development of intrafolial folds by the flattening of larger structures.
B. The occurrence of intrafolial folds between 'undeformed' layers.

tional strain produces a great diversity in fold style.

As more detailed work is collated on mobile belts it is becoming apparent that the various attributes of folds collectively referred to as 'style' may change considerably over short distances. Tobisch et al. (1970) have examined a 450 km area of Moinian and Laxfordian rocks in the Scottish Highlands and have shown that minor folds show a wide range of style and orientation over distances as small as 5 km.

For these reasons the writer feels that any correlation of folds and fold sequences between the Onseepkans area and other areas in the Namaqua belt is not possible at this stage and therefore such a correlation has not been attempted.

Plate 54. Rootless fold in biotite schist horizon intercalated with quartzo-feldspathic gneisses. Orangefall Farm.

CHAPTER VII

SUMMARY

The rocks in the Onseepkans area consist of a series of layered gneisses (the Onseepkans sequence), intrusive granitoids and mafic bodies. No evidence was found to indicate that the layered gneisses represent a cover sequence and there is no evidence that any of the granitoids constitute a basement. Only a small percentage of the layered gneisses can be positively identified as metasediments. Quartzites are virtually restricted to the area south of the Pofadder Lineament while the Beenbreek megacrystic granite and the Eendorn granite, which is its equivalent in the Warmbad area, are restricted to the area north of the Lineament. Rocks of pelitic composition are represented by quartz-muscovite schist (Kambreek formation) south of the Pofadder Lineament and by cordierite-garnet gneiss (Jerusalem formation) in the north.

The earliest detectable metamorphic event in the area took place under anhydrous granulite-grade conditions approximately contemporaneously with the intrusion of a series of charnockitic suite rocks (Stolzenfels norite and enderbite formation). Parts of these bodies are now represented by mafic granolites and these grade imperceptibly into isotropic rocks. This metamorphic event (the Kumian) produced garnet and sillimanite in rocks of pelitic composition. Later retrograde events have largely eradicated the earlier parageneses and the effects of Kumian metamorphism in other formations are not well known. Xenoliths in the Beenbreek megacrystic granite preserve Kumian parageneses in some of the mafic granolites and in the Jerusalem formation. The P-T conditions of this event were 8-9 kb at 800° - 860°C. Charnockitic suite rocks, orthopyroxene-bearing metamorphites and garnetiferous rocks of pelitic composition are absent from that part of the area south of the Pofadder Lineament.

A later metamorphic event (the Velloorian) was associated with the widespread development of migmatites, the intrusion of the Beenbreek megacrystic granite and the Naros granitoid. The Velloorian is divisible into a pre-Beenbreek phase (M_1) and a post-Beenbreek phase (M_2). Both phases appear to have developed under high-grade metamorphic conditions defined by 'quartz + muscovite out'. The Velloorian resulted in the retrogression of pyroxenes to hornblende in the mafic granolites and the marginal areas of the norites. In the Jerusalem formation extensive rimming of garnet by cordierite is thought to be due to this event. South of the Pofadder Lineament medium-grade metamorphites defined by 'quartz + muscovite in' are found in rocks of suitable composition. Cordierite only is found in these rocks and it appears to be a prograde mineral unlike the association found in the northern block. P-T conditions of this event in the northern block were 650^{o} - 700^{o} between 5 and 7 kb and in the southern block 650^{o} C between 3,5 and 5 kb.

The very pronounced change in metamorphic parageneses across the Pofadder Lineament is compatible with the interpretation of the southern block as being a preserved upper-crustal segment.

The major crustal break now represented by mylonites of the Pofadder Lineament forms the core of a shear zone (Pofadder ZAHNCAFS) that varies in width from 20 - 40 km. In the zone of reorientation adjacent to the mylonite in the northern block a series of northeast trending, en-echelon doubly plunging folds have been formed (D_5) as a result of destral displacement in the shear zone. Some of these folds are over 30 km long and at the margin of the ZAHNCAFS they pass into open dome and basin structures. A component of pure shear in the development of the ZAHNCAFS is thought to have been responsible for the refolding of the D_5 structures to define a younger northwest trending fold set (D_6). Neither D_5 or D_6 folds are present south of the mylonite belt and this may be due to a strain shadow effect connected with the presence, farther west, of the large rigid mass forming the Vioolsdrif complex. A minimum displacement of 85 km is indicated if the development of the shear zone involved a 40% pure shear component.

The lack of retrograde effects in the zone of reorientation and the partial anatexis connected with some minor shear zones suggests that the Pofadder ZAHNCAFS formed under medium to high-grade metamorphic conditions. The quartz fabric in the mylonites shows that these rocks have been formed by dislocation and recovery - a dynamic process typical of hot working.

Two earlier sets of macroscopic folds have been identified (D_3 and D_4) and in the eastern sub-area, where the effects of the Pofadder ZAHNCAFS are not seen, it is apparent that these earlier folds show a wide range in trend. The D_3 Leopard synform in the western sub-area is the only major fold with a fairly well developed axial plane cleavage (s_3) but fold hinge linear fabrics are a feature of many of the major structures. Two earlier periods of deformation are identified on the basis of foliated xenoliths found in the Beenbreek megacrystic (D_1) and in the Naros granitoid (D_2). No folds formed during these periods of deformation have been recognised. The pre-Beenbreek foliation (s_1) is well developed in the area as a whole and together with s_3 probably forms the regional foliation.

A quantitative study by harmonic and t_1/α analyses of D_3 and D_4 mesoscopic folds has shown that the fold style is largely dependent on the lithologies defining the folds. Competent horizons have a class 1C form and indicate a formation by buckling with high components of tangential longitudinal strain or sometimes flexural slip. Incompetent horizons are usually those richer in biotite and they have a class 3 form. The overall effect is class 2 (similar fold) but the quantitative studies have shown that these are not true 'shear' folds. It has also been shown that different generations of structures do not have a unique fold style but the lithologies which produce certain patterns for D_3 folds will produce the same patterns for the younger, D_4, folds.

Finite strain estimates on D_3 and D_4 folds show total shortening in the fold profile plane varying from a mean of 1:9 for D_3 folds to 1:6 for D_4 folds. However, considerable variations exist for each major structure examined. This gives an estimate of the shortening suffered by the layered rocks in the Onseepkans area. Such studies can eventually lead to the determination of the shortening across the Namaqua belt.

Five out of six age dates (Appendix 1) suggest that the '1000 Ma event' recognised elsewhere in the Namaqua - Natal Mobile belt may also be recognised in the Onseepkans area. However, this figure may indicate the closing of the system as age dates appear to change south of the Pofadder Lineament where high-level rocks are preserved.

APPENDIX I

Geochronological data on rocks from the Onseepkans area

P.R.U. No.	Locality	Farm Nama	Rock Type	Isotopic Values			Calculated Ages			N207/ 235	N206/ 238
				Pb 206/204	Pb 207/204	Pb 208/204	Pb 207/206	Pb 207/235	Pb U238		
103	28°40'S	Velloors-drift	Orangefall biotite gneiss	3,651	259,6	325	1142	866	764	1,3214	0,1246
104	28°37'S 19°15'E	Velloors-drift	Orangefall biotite gneiss	80,000	22,17	58,45	1720	1384	1042	2,4849	0,1737
105	28°44'S 19°24'E	Been-breek	Beenbreek megacrystic granite	2206,5	190,8	156,05	1218	1016	928	1,6869	0,1532
106	28°29'S 19°31'E	Onder-matje	Naros granitoid	4633,92	354,26	504,73	1042	868	802	1,3235	0,1311
112	29°40'S 19°09'E	Keimas	Nautsis alkali-feldspar granite	447,21	45,86	95,75	952	402	314	0,4773	0,04939
117	28°24'S 19°37'E	Stolzen-fels	Stolzenfels norite	3559,99	284,34	4569,60	1110	524	404	0,6651	0,6402

Dr. A.J. Burger, National Physical Research Laboratory of the South African Council for Scientific and Industrial Research, has determined minimum U-Pb ages by single zircons in rocks from the Onseepkans area. These results are preliminary.

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